THE SPRING LAKE ANTICYCLONE

Its inducement on the atmospheric and water circulations

by
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ABSTRACT

An air-water interaction study was performed near Waukegan, Illinois, to determine the effect of the cold waters of Lake Michigan on warm air flowing over them. Nine days in late May and early June were chosen to analyze the physical nature of the interaction. Many times during spring and early summer, a cold temperate lake produces intense inversions that limit vertical turbulent energy transfers. Within such an inversion there often develops a shallow overwater mesoscale high pressure system, the lake anticyclone.

The observations of the cold lake's effect on the atmosphere were taken on an east-west line through Waukegan for a distance 5 km inland (west) and ranging as far as 40 km over the lake (east). Hourly wind soundings to a height of approximately 1500 m were made and air and dew point temperature profiles were measured to 300 m by wiresonde from the R/V (research vessel) MYSIS. Additional ship measurements of water temperature and of air temperature and wind at 11 m were made to obtain Richardson number values in the inversion and to show air-water interaction effects. Pilot balloon observations at the shoreline and 5 km inland were analyzed together with balloon soundings over the water to depict the diurnal development of the lake's effect on the gradient wind near the shoreline.

The low level wind near the lake was always altered by the prevailing regional wind above the inversion. This lake influence was observed on every sunny day. The major orthogonal gradient winds and their interaction with the lake anticyclone were compared with results from a numerical model (Estoque 1962) of "sea effect." The most

dramatic effect occurred when a warm offshore wind moved out over a developing inversion above the lake. As the air approached the shoreline it experienced a pronounced upward component of motion, followed by strong subsidence as it moved out over the lake.

Parallel-to-shore wind data from Waukegan demonstrated increased vertical motions from the balancing forces of the northerly gradient wind and the lake high over the reverse case of the southerly gradient wind and the lake high. Results under onshore gradient winds, while least dramatic in vertical motion induced by this lake, demonstrated low level cloud suppression for several tens of kilometers over land downwind.

The direct air-water coupling effects imposed by wind stress under differing gradient wind regimes aided or hindered the thermal current circulation of the water during spring. Edge-warming, caused by the penetration of solar radiation into shallow water and thermal energy from river runoff, left the central deeper waters much colder well into June. The resulting cyclonic water transport in Lake Michigan was the recipient of minimized turbulent energy transfers down through the stable lake anticyclone. The lake high remained a positive feature until the air-water stability ($T_{air} > T_{water}$) deteriorated later in the summer.

INTRODUCTION AND LITERATURE REVIEW

Lake breezes, sea breezes, and the Great Lakes

The role of the Great Lakes in various energy transfers with the atmosphere is just beginning to be understood. The geographical position of these large bodies of water makes it possible for them to encounter the extreme contrast of the seasonal weather patterns. Warm tropic-like air may reach the Great Lakes early in the spring when the water is still close to maximum density. During the fall, early outbreaks of cold arctic air sweep down over the Canadian Plains to the lakes which still retain a large portion of the thermal energy that was acquired during the summer. With these temperature differences extreme instabilities occur. These two periods produce extremes in over-water stabilities which are unequalled by most bodies of water. Stabilities with lapse rates as strong as 25°C/100 m have been reported by Bellaire (1965) in early May. Instabilities equally as great have been measured by airborne instruments in December and January (Lenschow 1965).

Instability produces an increased incidence of cloudiness (Davidson 1967) and convective precipitation (Changnon 1967) over the entire Great Lakes region. Energy transfers are maximal during the strong convective addition to the circulations of autumn (Strong 1968). Recent studies of "lake effect" snowfalls have been reported by McVehil and Peace (1965) and Thomas (1964). The Lakes' interaction with, and influence on, the larger scale weather patterns are greatest during these several months of instability when dramatic heat and momentum exchange persist.

Typical of the fall and winter season is a low pressure trough associated with warmer water feeding energy to a cold and extremely unstable atmosphere (Petterssen and Calabrese 1959). Although the fall and winter lake effect weather is a product of each individual lake, the extreme vertical extent to which the exchange is affected integrates each contribution into a larger scale phenomenon which may influence a major portion of the weather over the eastern United States.

Quite the opposite condition exists during the spring. Strong stabilities which result as the cold water temperatures respond sluggishly to the warming air temperatures cause a reduction in air-water interactions. Contrary to the fall low pressure effect of the Great Lakes, high pressure dominates, and is a function of the springtime inversion.

When a strongly stable atmosphere is examined by some stability parameter (e.g. Richardson number), there is the indication that heat energy transfer becomes negligible as the warm air is buoyed above the air which is being cooled by conduction to the cold water surface below. Momentum transfer in the inversion also becomes drastically reduced. McVehil (1964) and Lettau et al. (1967) have presented results of some of the few investigations made under the positive Richardson number (Ri) regime of inversions.

During stable days in May and June Ri numbers within the inversion typically vary from 0 to +1. Occasionally much higher values can be found within the lower 100 meters over the lake surface. Some authors argue that a flow near the ground becomes laminar at Ri = +0.3 (Portman et al. 1962) while others put this value closer to +1.0 (Richardson 1920). If the flow within the inversion approaches a laminar state, K_H becomes

zero and K_M becomes very small. The ratio K_H/K_M becomes infinitely small. It is necessary to treat the lake as having a nearly isolated atmosphere, which during the spring months is greatly restrictive to the the transfer of momentum through the interface from the prevailing gradient wind aloft.

When warm air moves slowly from the land out over the water, and solar radiation provides a large land-water temperature contrast, the classical lake breeze results. A numerical lake breeze model for the Great Lakes has been developed by Moroz (1967) for zero gradient wind conditions. Estoque (1961) has produced a numerical model for the sea breeze into which he later imposed a geostrophic wind (1962). It will be shown that with a gradient wind of 5 m/sec a Great Lake is of sufficient size to be modeled as an ocean, and Estoque's results become "valid" also for the lake.

During the spring and early summer, the lake circulation will be largely a function of the dynamics of the cooler air mass lying within the restrictive inversion. Any persistent vertical structure in this "blanket" of air may induce, through divergent or convergent horizontal wind stress, sinking or upwelling water. Moroz (1965) has suggested that the lake breeze's effect on the surface water might be such as to break down the driving temperature differential. It was subsequently suggested by Rodgers (1966) that an upwelling persists during spring in the central region of a Great Lake. Such upwelling might be partially maintained by atmospheric subsidence that would induce a local barometric high pressure region, resulting in low level divergent wind stress at the water surface in the center of the lake. This divergent water movement toward the lakeshore, in conjunction with the lake-edge warming

(Ayers and Strong 1967), may produce the springtime "thermal bar." This principle has also been discussed by Rodgers and Anderson (1963) and Rodgers (1965).

The thermal bar is a circulation phenomenon peculiar to temperate fresh water lakes. The key factor involved is that water is most dense when it is at 4°C. As the Great Lakes warm in the spring from below 4°C to above 4°C there is a period when the deep, mid-lake water is below 4°C and the shallow nearshore waters are well above this critical temperature. Due to sinking along the 4° isotherm, a semi-isolation is briefly (several weeks) established between the waters on either side of the region of most dense water. The isolation tends to persist until the mid-lake and colder surface water warms to 4°C. Such a restrictive mechanism could provide colder than maximum density water at the surface of the central lake until the thermal bar weakens. Maximum over-water stability is provided just before the thermal bar vanishes toward the end of May. Some of these facets are covered in an initial investigation of Lake Michigan climatology by Ayers (1965) and later updated by Ayers and Strong (1967).

Extension of observations of the lake effected atmosphere over the water has provided data that until recently have been extremely limited. Observations of sea breezes in the tropics by Donn, Milic, and Brilliant (1956), Staley (1957) in the Pacific Northwest, Malkus, Bunker, and McCasland (1951) in New England, and Wexler (1946) in Europe, Africa, and England have consisted almost exclusively of land measurements. A few observations by Frizzola and Fisher (1963), Angell and Pack (1965), and Craig, Katz, and Harney (1945) have provided nearly all that is known of the over-sea temperature, vapor pressure, and air movement in

sea breeze conditions. Reports of local winds on the Great Lakes have recently included more over-water measurements taken by commercial ships and research vessels. Moroz (1965, 1967) utilized shipboard measurements as did Munn and Richards (1967) in studies of the lake breeze. The author (Strong 1967) reported a wiresonde[1] profile made during a strong lake breeze development into a weak offshore gradient wind.

In all these lake and sea breeze studies the gradient wind was nearly absent and consequently neglected. In fact, the ideal situation for measurement came when a weak pressure gradient existed. To incorporate the gradient wind into lake breeze studies it was necessary to measure the effect on the prevailing gradient wind as it interacts with the masked lake breeze. The term "lake effect", consequently, becomes a more general description of the altered atmospheric circulation over a large lake. The terms "lake breeze" and "snowbelt" are, then, specific examples of lake effect.

Recent studies by Hewson et al. (1960, 1963) and Lyons (1967) have begun to incorporate the effect of the Great Lakes on the diffusion of atmospheric pollutants. As man continues to concentrate around these five bodies of water the foul air problem grows. The lake breeze has been shown to concentrate aerosols into a convergence zone inland[2] near the lake shore whence they rise above the lake air and may return over the lake in air currents aloft (Changnon 1968). The convergence zone acts as a line of giant smoke stacks at the inland boundary of the lake effect.

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[1]A teathered balloon was used to lift dry and wet bulb thermisters above the lake surface.

[2]The lake breeze front - or zone that separates warmer air inland from the cool lake air.

While man pollutes the atmosphere he routes the major portion of his wastes back into the lakes as "treated sewage." The portion that enters the lake during spring becomes captive in its environment, and the alongshore currents carry the problem elsewhere. Whose immediate problem this effluent becomes depends largely upon the recent history of the atmospheric circulation. The lake effect and air-water coupling play roles in the disposal of the wastes of our increasing population.

If present population trends continue, the population density around the Great Lakes will double during the forty years from 1960 to 2000 (Doxiodis 1967).[3] Along with this influx seems to breed the environmental cancer of civilization: pollution. If the present rate of pollution increase continues we may not be able to breathe the air nor swim in the water by the year 2000. Solution to this problem lies in part with the understanding of mesoscale atmospheric circulations and their effects, not only on the large scale atmosphere but also in their interaction with the water.

Springtime Stability: The Inversion

By April the increasing solar radiation encourages modifying warmer air to advance northward into the Great Lake region. Since soil has a lower specific heat and thermal conductivity than water, the surface of the soil warms more rapidly than the surface of the water. Subsequently the air nearest the ground warms more rapidly than the air over the water and reaches a condition of neutral stability or instability soon after sunrise. Convective activity over the warm land generates fair weather cumulus clouds as its signature. On the lake solar energy is being absorbed by the water. However, the much higher transparency,

[3]From 36 million to 75 million in the states contiguous to the Great Lakes.

conductivity, and mobility of the water enables heat energy to be transferred to lower levels within the lake. The viscous nature of the water medium enables any subsequent turbulence in the surface water to mix heat to a deeper level within the lake. While the lake responds as a tremendous heat sink, the air adjacent to the water is not warmed, as over the land, but instead is cooled by conduction to the water. A neutral or unstable lapse rate is not established over the water, but instead, a cold stable environment is maintained. During a normal clear spring day a strong inversion develops over the lake. Cumulus development is absent, as convective processes are suppressed. The lake may be ringed by convective clouds, but they are rarely seen over the water. If advection carries them out above the lake atmosphere they are quickly dissipated. Lyons (1966) has shown graphic illustrations of cumulus suppression over Lake Michigan. He has shown the effect of the lake on a squall line as it advanced from the west; the thunderstorms died over the lake as if the cold lake were an inhibitor of convection.

One of the most dramatic observations of warm air modification over a cold lake occurred on 5 May 1964 (Bellaire 1965). Warm southerly winds preceding a cold front moved over cold Lake Michigan producing an inversion of 25.5°C (29.4°C air at 100 meters over 3.9°C surface water). the warmest water, close to shore, was 10.2°C; however, 25 kilometers lakeward the water was 5°C, and it reached maximum density 5 km further lakeward.

Reports of air-water stability ($T_{air} > T_{water}$) effecting the development of winds and waves have become more numerous in recent years. Ragotzkie (1962) has reported on temperate climate lakes which, observed from low level plane flights during spring, appeared smooth as if

ice-covered. Actual in situ observations later showed some of these surfaces to be in the liquid phase. Shallow insulating blankets of cold air were protecting the cold lakes from the surrounding atmospheric influences. The suppression of turbulent mixing shielded the lakes from the transfer of both atmospheric heat and momentum. Similar conditions, but on a larger scale, occur over the Great Lakes during spring and summer. It is the large areal extent of the Great Lakes producing these effects that make them quite unique. Greatest inversion development will be found just before the formation of the thermocline across the deepest portion of the lakes.

Other stability measurements have been reported by Brown (1953) on the Atlantic Ocean and on the Great Lakes by Lemire (1961), Strong and Bellaire (1965), Jacobs (1965), Richards, Dragert, and McIntyre (1966), and Cole (1967).

The ratio:

$$R = V_s/V_g$$

where V_s = wind measured at ship level

V_g = gradient wind

has been useful when comparing stability effects. As Richards et al. (1966) point out, this ratio decreases as stability increases. Under complete insulation conditions the ratio approaches zero and the Richardson number would become meaningless.

Attempts have been made at measuring vertical gradients under stable conditions overland. These efforts, unfortunately, have met with rather limited success. However, they represent what little we know of inversion characteristics. The primary concern has been with the Richardson parameter which through a finite difference approximation

becomes:

$$Ri = \frac{g \, \Delta z \, \Delta\Theta}{\overline{T} \, (\Delta V)^2}$$

where g = acceleration of gravity (m/sec)

z = height above surface (m)

Δz = thickness of the layer (m)

$\Delta\Theta$ = change in potential temperature upward through the layer (°K)

$\overline{T}$ = mean temperature of the layer (°K)

ΔV = change in wind velocity upward through the layer (m/sec).

The Deacon number (β) has also been utilized (McVehil 1964) where:

$$\beta = - \frac{\Delta \ln (\Delta V / \Delta z)}{\Delta \ln z} \quad .$$

Physically interpreted, the Deacon number is a measurement of the wind profile curvature which is affected by stability, while Ri characterizes the stability directly. Under inversion conditions all parameters making up the Ri number are positive. From observations made in Antarctica (Lettau, Dalrymple, and Wollaston 1967) during 1958, Ri values were reported as high as +0.5. The Richardson number profile increased with height above the snow surface to 400 cm. Above this level no information has been reported from the South Pole inversions by means other than rather gross rawinsonde measurements. Lettau et al. report that the inversion averages 600 meters in depth during the Antarctic winter night.

Additional work by McVehil (1964) deals with the turbulent exchange coefficients K_M and K_H. If u is the velocity of the horizontal flow, the momentum exchange coefficient is defined as:

$$K_M = \frac{\tau}{\rho \;\; \partial\Theta/\partial z}$$

and the coefficient of heat flux as:

$$K_H = -\frac{H}{Cp\ \partial\Theta/\partial z}$$

where τ = horizontal shearing stress

H = vertical heat flux

ρ = density of air

Cp = specific heat capacity of air at constant pressure

Θ = potential temperature.

McVehil treats the South Pole data and, in agreement with Lettau, shows that close to the surface there is a decrease with height in the ratio K_H/K_M during stable conditions. He further stipulates that this decrease becomes noticeable when $Ri \geq +0.08$ at 400 cm. While neutral conditions produce log-linear wind profiles, a Ri of +0.15 appears to signify an upper limit to this turbulent regime. McVehil finds further support for this relationship and threshold when comparing β and Ri. The relationship in positive Ri holds quite well until $Ri \geq 1/7$. Above +0.15 he can find "no consistent behaviour."

With this brief introduction of the poorly understood stable atmosphere it appears necessary to study the atmospheric stability condition that exists in the spring over the Great Lakes as a two part system: 1) The stable atmosphere that blankets the lake as a result of the cold water - The Lake Atmosphere, and 2) the large scale circulation, which represents a portion of the Midwest weather, and exists above the lake atmosphere. It is the interaction of these two media that will be investigated.

DYNAMICS OF THE LAKE BREEZE ENVIRONMENT

The lake high pressure system

With surface water colder than the overlying air, the persistence of a mesoscale high pressure phenomenon over the lake has been reported (Hall 1954; Bellaire 1965; Strong 1967). The lake high is commonly a separate feature of each of the Great Lakes. With several operating concurrently, general high pressure may, however, be induced over the entire region. Due to the low altitude to which the lake high operates, complete integration with the effect of other lakes is not often achieved. A true measure of the intensity of this small-scale pressure feature is difficult to achieve with present instrumentation. Sensors need to be maintained over the lake and in proximity with the lake shore. The pressure gradient normal to the shore may be quite strong, but very limited in its inland extent, making a measurement at an airport U. S. Weather Bureau station a few miles from shore not a representative index of the lake high.

The persistence of the lake high is seen on nearly every spring day when clouds are not too numerous. From past studies (Strong 1967) it has been generally seen that the maximum intensity of the lake high occurs between 1300E (1300 Eastern Standard Time) and 1600E barring any outside changes in the weather patterns. Maximum insolation appears to be essential for greatest intensity of the lake high. In addition, the water vapor- and haze-free atmosphere of the large scale high seems most conducive to a well developed lake breeze.

A study by Klein (1956-1957) shows the prevalent anticyclonic spring and early summer weather in the Great Lakes area. The tendency toward anticyclonogenesis can be noted from April through July when the water

temperature lags well behind the air temperature. As the thermocline develops and the spring overturn period is halted, more rapid heating of the surface water is possible. Only then will anticyclone inducement diminish.

The Estoque and Moroz models: Sea effect vs. lake effect

Estoque (1961) investigated the sea breeze by use of numerical modeling. His grid was centered at the shoreline and extended 200 kilometers over the water and 200 kilometers inland. From earlier observations of sea breezes he established a maximum height of the average sea breeze to be 2 kilometers. The lowest 50 meter layer of his model he hypothesized to be a region of constant flux - both heat and momentum. Above this sublayer he introduced a more realistic transitory region where eddy fluxes decreased with increasing height to vanish at 2 kilometers. The shoreline was straight and was represented by a normal through the model extending to infinity.

In the transitory region ($50 \text{ m} < z \leq 2000 \text{ m}$) the following linearized equations were deemed valid and were integrated (Estoque 1962):

The equations of horizontal motion and heating may be represented:

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + w\frac{\partial u}{\partial z} - fv = -\frac{RT}{p}\frac{\partial p}{\partial x} + \frac{\partial}{\partial z}\left(k\frac{\partial u}{\partial z}\right)$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + w\frac{\partial v}{\partial z} + fu = -\frac{RT}{p}\frac{\partial p}{\partial y} + \frac{\partial}{\partial z}\left(k\frac{\partial v}{\partial z}\right)$$

$$\frac{\partial \Theta}{\partial t} + u\frac{\partial \Theta}{\partial x} + w\frac{\partial \Theta}{\partial z} = \frac{\partial}{\partial z}\left(k\frac{\partial \Theta}{\partial z}\right)$$

where u, v, w are components of motion in the three orthogonal directions: x, y, and z.

t = time

f = Coriolis parameter

R = Universal gas constant

T = temperature

p = pressure

k = mixing coefficient

Cv = specific heat constant (volume constant)

Cp = specific heat constant (pressure constant)

The hydrostatic equation

$$\frac{\partial p}{\partial z} = - \frac{pg}{RT}$$

when combined with the continuity equation

$$\frac{\partial w}{\partial z} + \frac{\partial u}{\partial x} = - \frac{1}{\rho} \frac{\partial \rho}{\partial t}$$

and the heating equation to eliminate density (ρ) will yield his final equation:

$$p \frac{\partial^2 w}{\partial z^2} + \frac{\partial w}{\partial z} \frac{\partial p}{\partial z} = -p \frac{\partial}{\partial z} \left(\frac{\partial u}{\partial x}\right) + \frac{\partial}{\partial z} \left\{\frac{p}{\Theta} \frac{\partial}{\partial z} \left(k \frac{\partial \Theta}{\partial z}\right)\right\} - \frac{R}{Cp} \frac{\partial p}{\partial z} \left(\frac{\partial u}{\partial x}\right)$$

$$- \frac{Cv}{Cp} \frac{\partial u}{\partial z} \frac{\partial p}{\partial x}$$

He further simplified his final equation after finding the second term on the left and the second, third, and fourth terms on the right-hand side of the equation to be negligible in producing vertical velocity. The final equation then becomes

$$\frac{\partial^2 w}{\partial z^2} = - \frac{\partial}{\partial z} \left(\frac{\partial u}{\partial x}\right)$$

Below 50 meters

$$\frac{\partial}{\partial z} \left(k \frac{\partial u}{\partial z}\right) = \frac{\partial}{\partial z} \left(k \frac{\partial \Theta}{\partial z}\right) = 0$$

and the horizontal wind becomes:

$$U = (u^2 + v^2)^{1/2}$$

The mixing coefficient, k, is defined (Priestley 1959 and Estoque 1962) for the constant flux sublayer over water as:

$$k = [k_\circ(z - z_\circ)(1 + \alpha \ Ri)]^2 \frac{\partial u}{\partial z}$$

when $Ri \geq (Ri)_c$

and over land as:

$$k = \lambda z^2 \left(\frac{g}{T} \left|\frac{\partial \Theta}{\partial z}\right|\right)^{1/2} \qquad \text{when } Ri < (Ri)_c$$

if $(Ri)_c \equiv$ the free-forced convection transition = -0.03

$k_\circ$ = critical level

$\alpha \simeq -3$

$\lambda \simeq 0.9$

Above 50 meters k decreases linearly with height from the value prescribed at 50 meters to become zero at the 2 km top of the model. The driving force for his diurnal circulating sea breeze is a "close-to-realistic" sinusoidal temperature wave over the land. This forcing function is absent over the water, and at the shoreline has been reduced to half the land wave amplitude:

$$T = 283 + 10\sin(15t + 240°) \qquad x > 0 \text{ [land]}$$

$$T = 283 \qquad x < 0 \text{ [water]}$$

$$T = 1/2[T(x > 0) + T(x < 0)] \qquad x = 0 \text{ [shore]}$$

when t = hours after midnight

The $T_{air} - T_{water}$ maximum of 10°C occurs at 1400 local time after 0800 LCT produced the initial time when $T_{air} = T_{water}$.

Estoque (1962) produced a more realistic model of a perturbation on the sea breeze by incorporating a geostrophic wind into the upper layer of his earlier model. The resulting motions typify what may be called the "sea effect." In his original work on the sea breeze model he

directed a 5 m/sec prevailing wind in four directions:

1. Offshore
2. Onshore
3. Parallel to shore into the model
4. Parallel to shore out of the model

In comparing these results one finds relatively constant "sea effect" beyond 32 kilometers over the water. His working numerical model extended to 200 kilometers on each side of the shore. By taking a 180° reversal of the x axis of his offshore gradient wind results and joining it to the onshore results one can obtain a "lake effect" model for a wind blowing across a large lake. This limit in apparent shoreline influence beyond 32 km demonstrates that spring lake effect on the Great Lakes should have a sea effect characteristic when a 5 m/sec geostrophic flow is present.

For the parallel to shore geostrophic flow, a similar inverting of one of Estoque's gradient flow results with the opposing flow will provide a composite prevailing alongshore lake effect wind model. Two of these paired models are presented in Figures 1 and 2. For Lake Michigan, these prevailing winds would be westerly and northerly, respectively. (The coupled lake width of approximately 140 kilometers is in close agreement with Lake Michigan's east-west width at Waukegan, 126 km.)

The most dramatic effect occurs at the upwind shore under normal-to-shore gradient winds (Fig. 1). A northerly west-shore (Fig. 2) wind or a southerly east-shore wind (a reverse of Fig. 2) causes considerable vertical interaction with the lake breeze.

For geostrophic winds less than 5 m/sec a transition must be made from Estoque's model toward a zero gradient wind lake breeze model

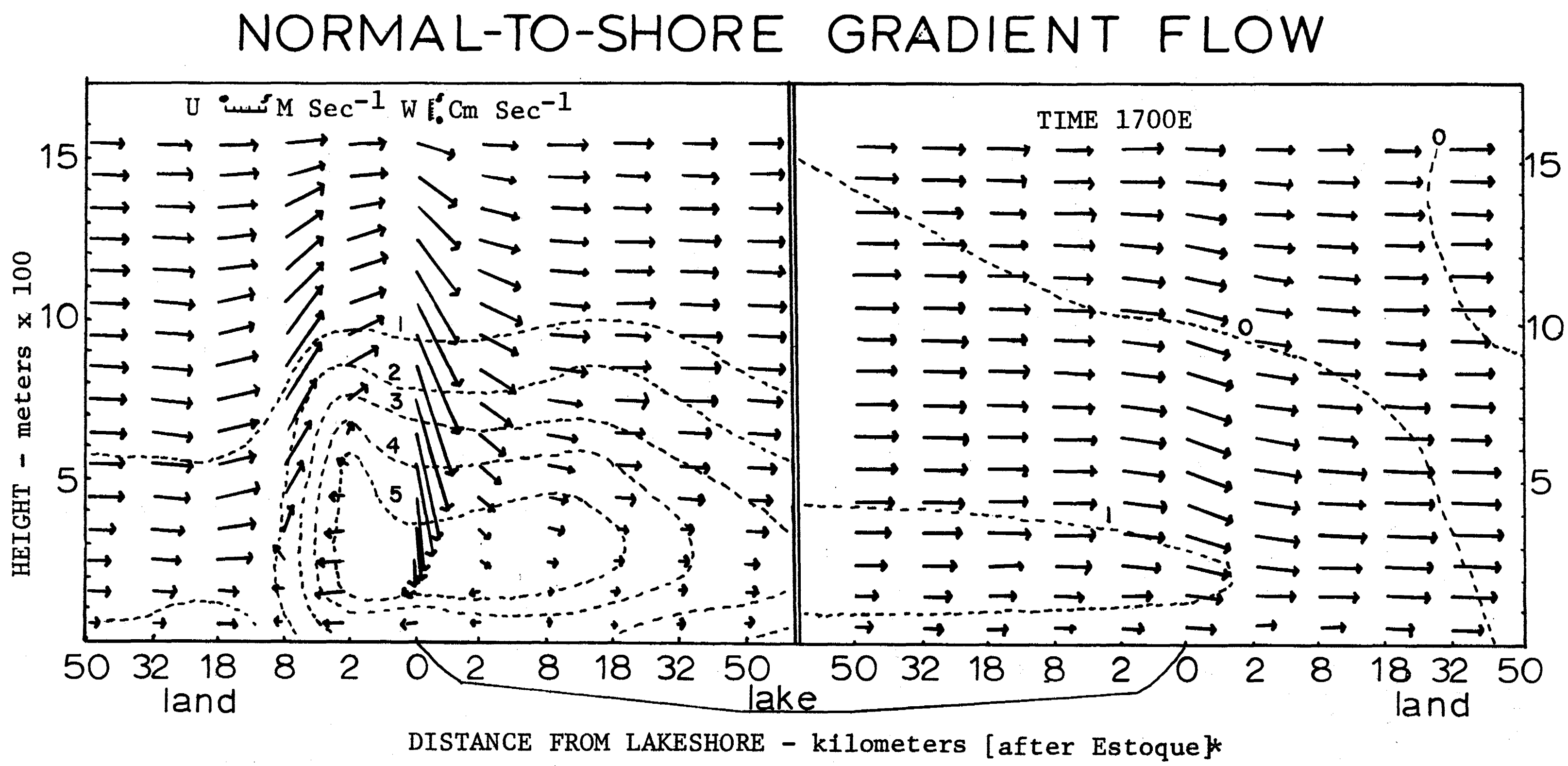

FIG. 1. Normal-to-shore 5 m/sec gradient wind flow over Lake Michigan [after Estoque]. Vectors represent normal-to-shore and vertical circulation. Dashed lines represent flow (m/sec) into figure. The flow, in the normal sense, would depict a westerly gradient wind.

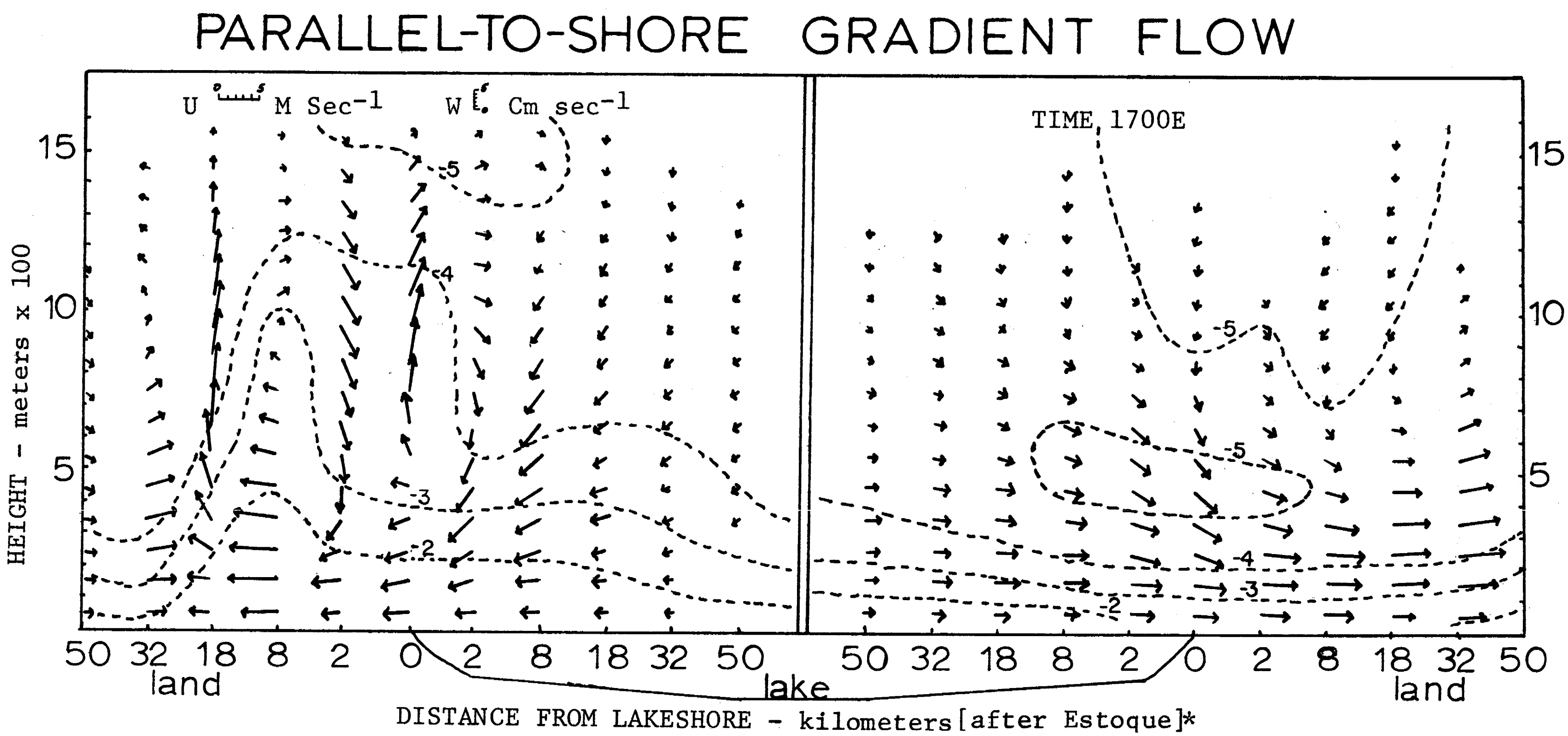

FIG. 2. Parallel-to-shore 5 m/sec gradient wind flow over Lake Michigan [after Estoque]. Vectors represent normal-to-shore and vertical circulation. Dashed lines represent flow (m/sec) into figure. The flow, in the normal sense, would depict a northerly gradient wind.

presented by Moroz (1967) for a cold lake. Estoque has indicated that his numerical results are in reasonably good agreement with observations. It would then appear that the lake effects of the Great Lakes are essentially equivalent to "sea effects" except at low gradient wind speeds.

The difference of the overlying atmosphere of a cold lake and the larger oceanic counterpart is most pronounced when the gradient wind is absent. While the initial development of both Moroz's (1965) and Estoque's (1962) lake and sea breeze is similar, the sea breeze develops a more intense push inland during the afternoon. This stronger feature may be attributed to the larger expanse of the water involved with the "pure" sea breeze. The "pure" lake breeze is limited by the double cell system that involves opposite lake shores. The divergent flow near the center of the lake may extend subsidence further lakeward than in the single-celled sea breeze while limiting horizontal motions near the lake's center. Use of the revised Estoque model will be employed throughout this study, as the gradient wind was close to or greater than 5 m/sec during the observation period.

THE OBSERVATIONAL STUDY

The nearshore observation program

A late spring period was chosen to investigate the lake anticyclone. This period was desirable because the greatest temperature contrast between water and air has been shown by Church (1945) to occur at this time. The period 22 May to 8 June 1967 was chosen, with ideal (less than 50% cloud cover) lake effect conditions actually occurring on nine of these days: 24, 25, 26 May and 1, 2, 3, 4, 5, 8 June. The main concentration of the study was along a nearly east-west line through Waukegan, Illinois, extending 5 kilometers inland and to a maximum of 40 kilometers lakeward. A map is presented in Figure 3. The shoreline runs nearly north-south at Waukegan.

Standard weather observations were obtained by our personnel on land at the shore (Waukegan Yacht Club) and 5 kilometers inland (WKRS radio station) in an open field. These observations were made generally on the hour, and they were concurrent with our pilot balloon (pibal) soundings. These upper wind measurements were initiated along the east-west line using single theodolite sitings on land. At sea a sextant-azimuth combination instrument, as described by Moroz (1965), was employed. Measurement of the 30 gram (pibal) balloon's flight was made every 15 seconds during its flight. Cloud and smoke-plume photographs were obtained whenever the situation warranted them.

Over-water observations were made from the R/V (research vessel) MYSIS. A meteorological recorder on the ship maintained a minute by minute record of the variables shown in Table 1 whenever the ship was at sea. The major efforts of the MYSIS personnel, however, were in obtaining

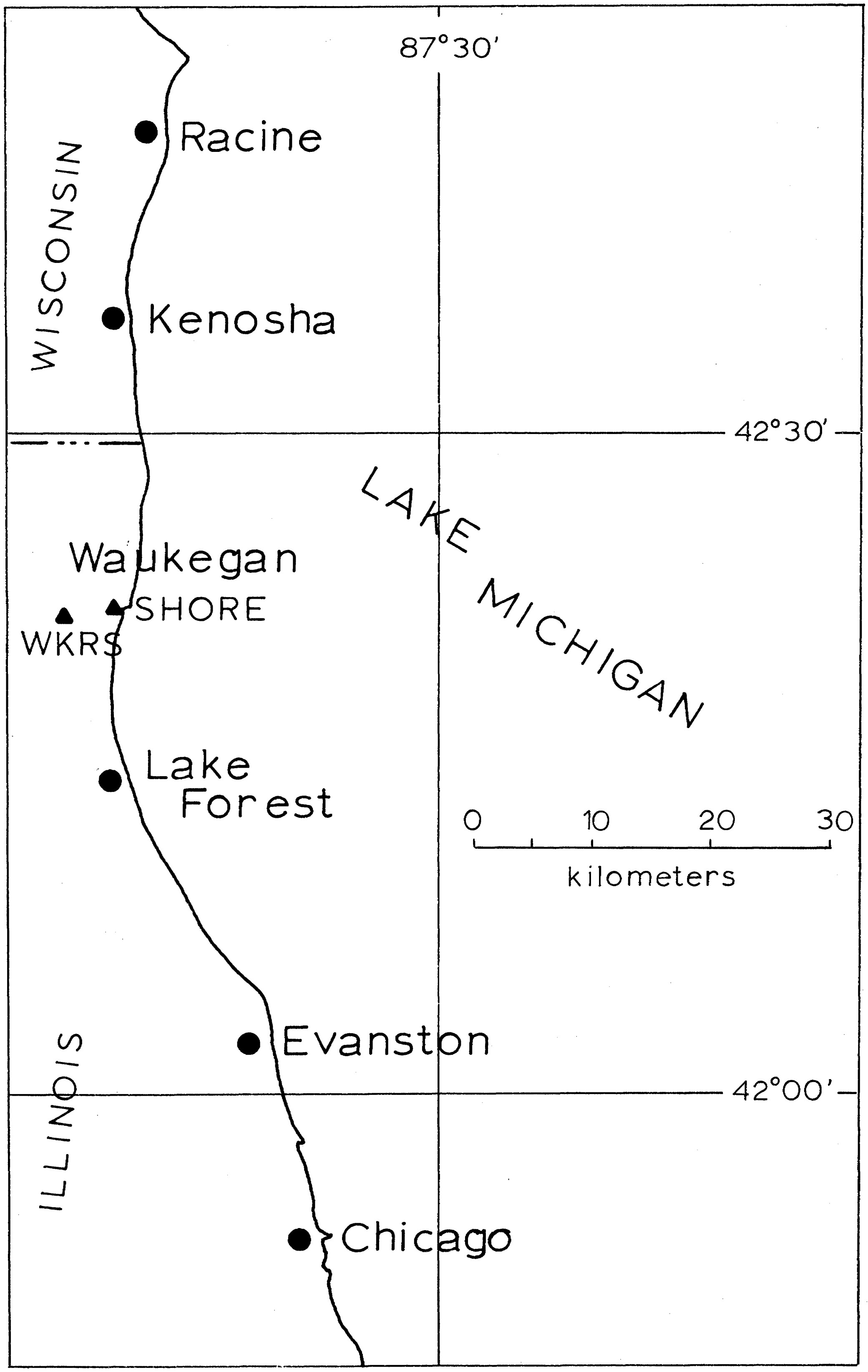

FIG. 3. Observation site along the west shore of Lake Michigan.

TABLE 1. Air-water observations from the R/V MYSIS and shore.

MYSIS - 0-40 km off Waukegan		
Air temperature	10 m	above water surface
Wind	10 m	
Air temperature	4.5 m	
Dew point	4 m	
Intake water temperature	-1.2 m	
Upper winds - Pibal		
Marine weather observations		
SHORE and 5 KM INLAND		
Surface weather observations		
Upper winds - Pibal		

hourly pibal soundings at various distances offshore and 0-300 meter wiresonde profiles through the inversion. These two operations provided wind profiles from 0-1500 meters and air and dew point temperatures in the lowest 300 meters. Radio communication was maintained between the MYSIS and shore facility to assist in coordinating operations as conditions changed.

The characteristic Richardson number was calculated for all the days of the Waukegan study. The height intervals (Δz) considered were: 0-50 meters and 0-10 meters. The top of the inversion has been found to be generally 50 to 100 meters over the lake so that the first interval Richardson number represents some "average" indication of inversion strength. The second interval, 0-10 meters, was within reach of the shipboard sensors. A summary of the daily results is presented in Table 2. The values ranged from a single null value for the 0-10 meter layer

TABLE 2. Average inversion and ship layer Richardson number.

Date	0-50 meters*	0-10 meters
24 May 1967	0.98	0
25 May	1.55	3.79
26 May	1.30	1.32
1 June	0.34	1.48
2 June	1.14	0.30
3 June	0.64	3.18
4 June	0.64	0.64
5 June	0.84	1.34
8 June	0.86	4.00
25 May 1966 (Strong 1967)	1.19	0.76
5 May 1964 (Bellaire 1965)	0.34	11.8

* Approximate time for value is 1300E.

to a maximum of 4.0 for the same layer. The null value occurred under a cooler gradient wind parallel to shore (northerly). The highest Ri value of the period occurred on 8 June when a warm brisk offshore (west-southwesterly) wind prevailed. Other values of Ri for previous Lake Michigan studies have been included for comparison. In all cases the lowest measurement in the layer was that at the water surface or the estimated wind necessary for producing the observed sea state (smooth = 0, ripples = 2, wavelets = 5, whitecaps ≥ 8 knots).[4] The 50 meter wind for 5 May 1964 was estimated from the wiresonde angle at this level. Because the wind shear through the inversion is large, the error introduced by these estimations is not too significant. A micrometeorological study is direly needed for these measurements.

[4]On three days of the period the water surface was smooth.

Anciliary observations

Large amounts of data were gathered from other sources in the proximity of Lake Michigan. Mr. Walter Lyons, of the University of Chicago, maintained an admirable record of pibal observations and cloud photography during the study period at Chicago, approximately 50 kilometers to the south of Waukegan. Mr. Lyons made several investigatory plane flights during a few days for cloud photographs and air temperature measurements.

Daily satellite pictures of the Great Lakes region were obtained from the National Environmental Satellite Center's ESSA 3 and ESSA 5 satellites. These weather satellites provided pictures of cloudiness at about 1400E each day during the Waukegan program. Similar pictures have been used in previous studies to show lake and sea effect activity (Parmenter 1967). They are presented in the Appendix - Figures A1-A9.

The R/V INLAND SEAS spent several days within the period in the southern basin of Lake Michigan. Her visit provided additional over-water weather observations that could be incorporated to provide additional checks for this study.

Further data incorporation became possible through a meteorological installation maintained by Brookhaven National Laboratory at a future nuclear powerplant site 10 kilometers south of Benton Harbor, Michigan, (location shown in Fig. A10). The American Electric Power Service Corporation, through the Brookhaven Laboratory, provided 16 and 67 meter wind data within 1 kilometer of the beach. Another recording system at the power plant site, maintained by the Great Lakes Research Division, provided alongshore currents and shoreline winds at the base of a lakeshore bluff.

Other data were available from the 42 Weather Bureau stations within 240 kilometers of Lake Michigan, 11 of which are within eight kilometers of the lake. Weather observations from commercial ships and regular car ferry operations produced the remainder of the data incorporated into the study.

Daily regional weather analyses

During the spring season the mesoscale weather analysis (use of all data available) quite often portrays a much different picture than the larger scale U. S. Weather Bureau analysis. Figures A10-A19, in the Appendix, show the 1300E mesoscale analyses for the nine days of the period of this study. While 45 stations (in addition to ship reports and the Waukegan observations) were used to make the analyses, for the sake of clarity only 25 stations plus ship reports and Waukegan weather have been presented on these maps. Figure A10 shows the 25 stations represented. The 20 stations used in the analyses but not shown on the maps were:

Illinois	Indiana	Michigan	Ohio	Wisconsin
Bradford	Bunker Hill AFB (Kokomo)	Alpena	Findlay	Lone Rock
Chanute AFB (Rantoul)		Battle Creek	Toledo	Oshkosh
Joliet		Flint		Rhinelander
Moline		Grand Rapids		
		Houghton		
		Jackson		
		Oscoda		
		Saginaw		
		Ypsilanti		

The pressure patterns are shown with a one-millibar spacing. Air (upper figure) and dew point (lower figure) temperatures are included with the wind and cloud cover for each station. Ship observations have water temperature substituted for dew point temperature. Due to the closeness of the three Chicago airport weather stations, they are

indicated in Figure A10 by the airport name only: Meigs, O'Hare, and Midway.

Although the most intense lake high observed during this study was of 2 millibars occurring on both 2 and 3 June, one as strong as 8 millibars was observed on 25 June 1965 (Olsson, Cole, and Hewson 1968).

The mesoscale weather observations from 25 May 1966 (Strong 1967) have been included (Figure A20) as supplemental data to the 1967 study. The observation site was identical for that day but lacked the onshore observations at Waukegan. These preliminary observations provide a comparison with conditions that are more characteristic of a pure lake breeze (weak pressure gradient). At Waukegan the offshore wind on 8 June (Figure A19) was nearly three times that of 25 May 1966. Restriction of the lake high pressure feature with increasing gradient wind is apparent. Table 3 groups the data by 1300E daily gradient wind orientation. Since the Waukegan shoreline is oriented nearly north-south an easterly gradient corresponds to an onshore flow, etc.

Observation days are presented as previously categorized in Table 2. If a gradient wind made an acute angle to a major wind division (N, E, S, or W) the smallest angle from a particular division designates the category into which the day was classed, e.g. south-southwesterly becomes classed southerly.

Better delineation of the lake high can be obtained by graphical subtraction of the large scale pressure pattern and the mesoscale gradient. The analyses of absolute vorticity show this lake effect equally well. Using all the 610 meter (2,000 feet) winds available at 1300E in the immediate vicinity of Lake Michigan, the Bellamy (1949) technique for calculating vorticity from pibal data was utilized. The

TABLE 3. Gradient wind classification for the study period.

Date	Wind* Direction	Speed (m/s)	Class
24 May 1967	350°	2	P-L
25 May	190	6	P-H
26 May	215	12	P-H
1 June	080	12	ON
2 June	100	5	ON
3 June	105	6.5	ON
4 June	190	8	P-H
5 June	195	7.5	P-H
8 June	270	20.5	OFF
25 May 1966	255**	3	OFF

KEY

ON = flow onshore.
OFF = flow offshore.
P-L = flow parellel to shore with lower pressure over water.
P-H = flow parallel to shore with higher pressure over water.

* 1300E wind 1.5 kilometers above the surface.
** 0.6 kilometers above the surface.

610 meter wind observation stations used are shown in Figure A21 of the Appendix. Figures A22-A31 illustrate absolute vertical vorticity as measured at 610 meters-433 meters above the mean level of Lake Michigan. The strength of negative vorticity correlates with divergence and higher pressure at this level. The inverse relation of gradient wind and the intensity of the lake high is apparent.

LAKE EFFECT WINDS UNDER DIFFERENT GRADIENT WIND REGIMES

Offshore gradient wind

Perhaps the most surprising results were obtained on 8 June 1967 under a 20.5 m/sec westerly wind (see Table 3). The strength of the gradient flow was instrumental in restricting the lake high pressure region over the lake surface. The interesting satellite cloud picture is presented in the Appendix (A9) and the mesoscale analysis in Figure A19.

A series of pibals were taken within one hour and twenty-five minutes of each other at the two shore stations and 5 kilometers nearly east of Waukegan from the MYSIS. These three soundings provide a cross section of horizontal wind vectors with height. These are presented in Figure 4. A vector pointing toward the right designates a westerly wind; upward is a southerly wind. It should be noted that this wind field differs from the portrayal of Estoque's model results in which the vertical wind is included. The soundings were as close to simultaneous as was possible. The time that each sounding was initiated has been indicated at the top of the sounding. Surface weather conditions at the shore throughout the afternoon changed relatively little. Cloud photographs at Chicago illustrate (Figure 5) this constancy.

At several locations along the lake's upwind shore smoke was observed to be carried nearly horizontally by the wind until over the lake. Jusy beyond the shoreline subsidence carried these plumes down nearly to the water surface in large wave-like motions. This feature at the water's edge was originally noticed at Waukegan about 1100E and persisted throughout the afternoon.

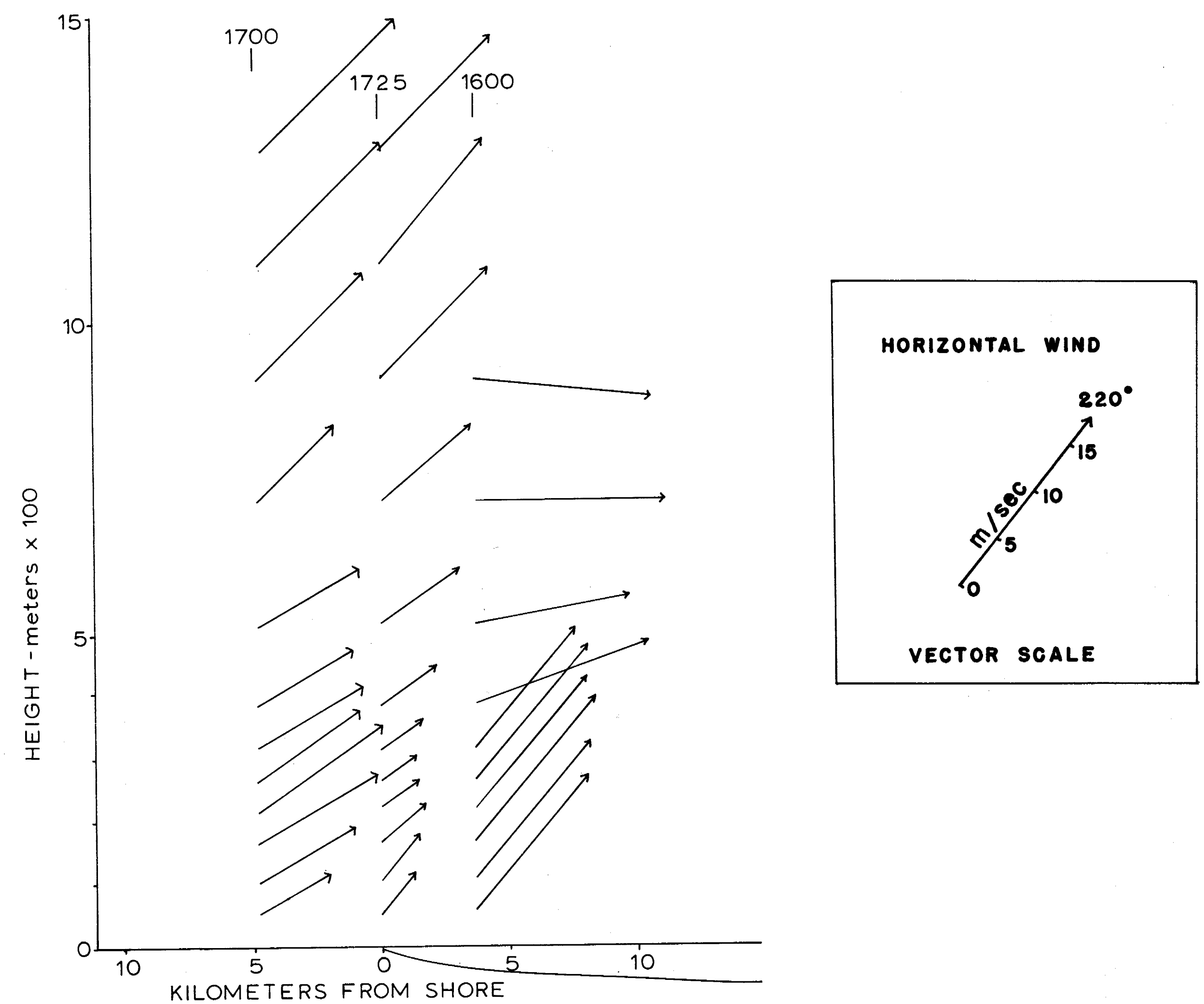

FIG. 4. Horizontal wind profile, 8 June 1967. Time of balloon release indicated above sounding.

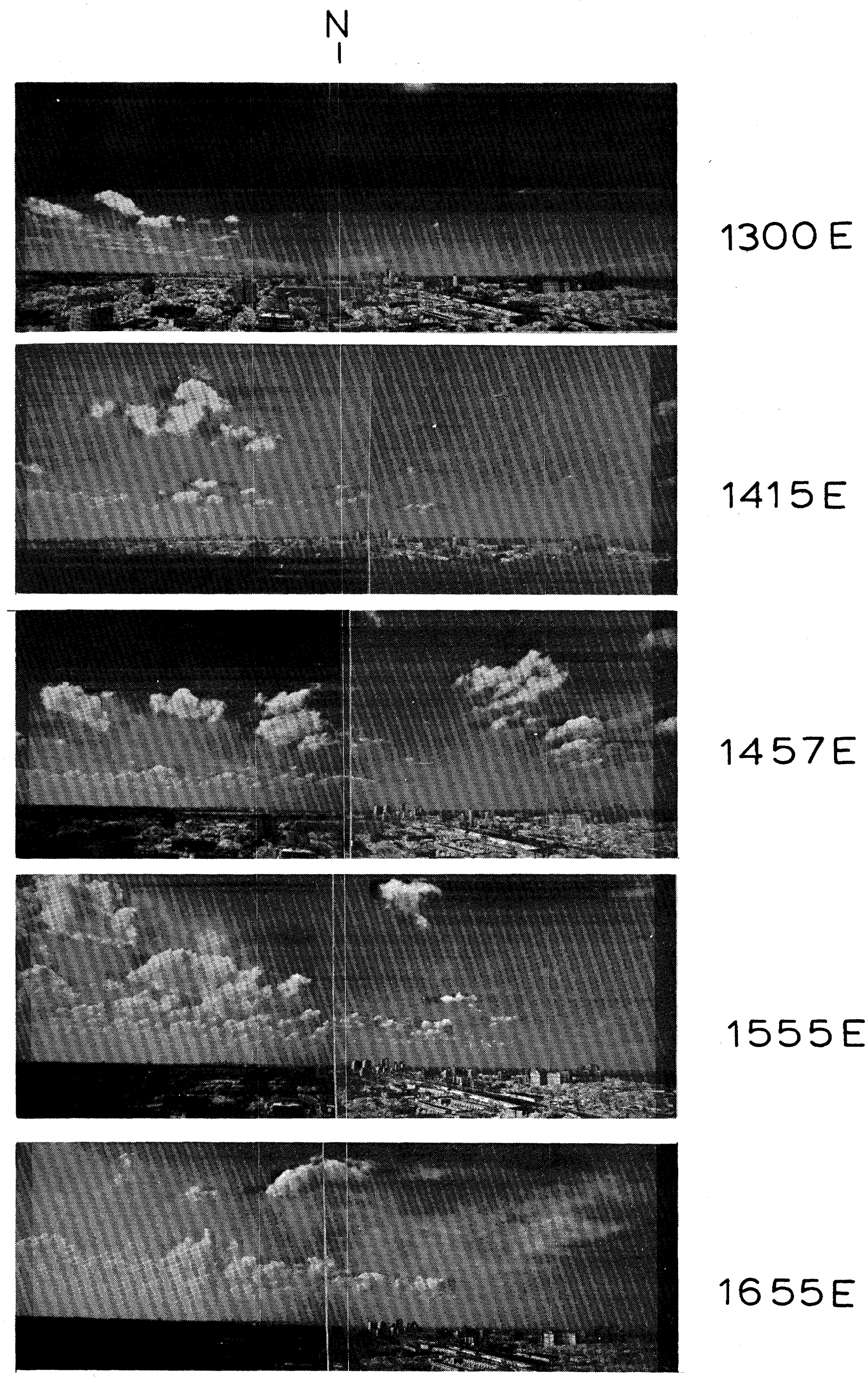

FIG. 5. Chicago cloud panoramas taken at the University of Chicago facing north (N) - 8 June 1967. The time for each panoramic is indicated beside the picture.

Air flow similarity was assumed north and south along the shoreline. Using this assumption, a second "shore" pibal sounding identical to the Waukegan sounding was then interpolated 5 kilometers to the north of Waukegan. This "measurement" was combined with the actual shore pibal and that from the MYSIS over the lake to give a triangular volume (prism) of the atmosphere that was investigated for horizontal divergence. Vertical velocity resulting from additive layers of calculated horizontal divergence was obtained as previously demonstrated by Bellamy (1949). A similar triangle was constructed with the WKRS and shore pibals. The resulting "soundings" of vertical velocity at the centroids of these triangles are provided in Table 4. The maximum vertical motion level appeared to be in the vicinity of 0.8 to 1 kilometer where the vertical velocity over the lake exceeded -100 cm/sec. The location of such extreme subsidence marks the horizontal limit of cumulus advection lakeward as well as significant lake upwelling that will be treated in a following chapter. The ascending current of air between WKRS and the shoreline must enhance any cumulus buildup in this inshore region. Cumulus turrets are present over the shoreline in Figure 5 at 1655E just prior to their dissipation over the lake.

The accurate measurement of single theodolite pibal soundings depends largely upon the atmosphere having no vertical motions. Pilot balloons are inflated to ascend at a rate of roughly 3 m/sec (600 feet/minute). A vertical wind velocity of -1 m/sec will slow the ascent rate by one-third, making the final calculation of vertical velocity lakeward subject to exaggeration of approximately the same order of magnitude. A similar argument for the convective region will show that the inland values are conservative. If the positive values of vertical velocity are

TABLE 4. Calculated vertical velocities for 8 June 1967 near the shoreline.

Height (meters)	W_{land} 2.2 Km inland (cm/sec)	W_{water} 2.2 Km lakeward (cm/sec)
1450	+33	---
1230	+35	---
1020	+36	---
814	+42	-101
619	+41	- 91
445	+35	- 74
345	+29	- 55
295	+25	- 48
245	+20	- 41
192	+14	- 33
138	+ 8	- 24
82	+ 4	- 15
28	+ 1	- 5

increased and the negative values decreased by 25-30% they should more accurately describe the character of this wave at the upwind shore. A correction of this amount would place the values of the opposed maximum velocities close to being equal (±60-70 cm/sec).

Presentation of the horizontal wind by U and V fields (eastward and northward components) in Figure 6 illustrate the shoreline "mountain" that results from the higher pressure associated with the cold air over the water. A diminishing of the lower level wind alongshore accompanied by a backing in direction indicate that the lake anticyclone has developed over the lake and that its effects have reached the shore. The pibal taken at 1600E from the MYSIS (4 km from shore) showed a wind of 220 degrees and 14 m/sec at 60 meters. Their surface weather observation, noting occasional white caps and a wind from 230 degrees at 6 m/sec at

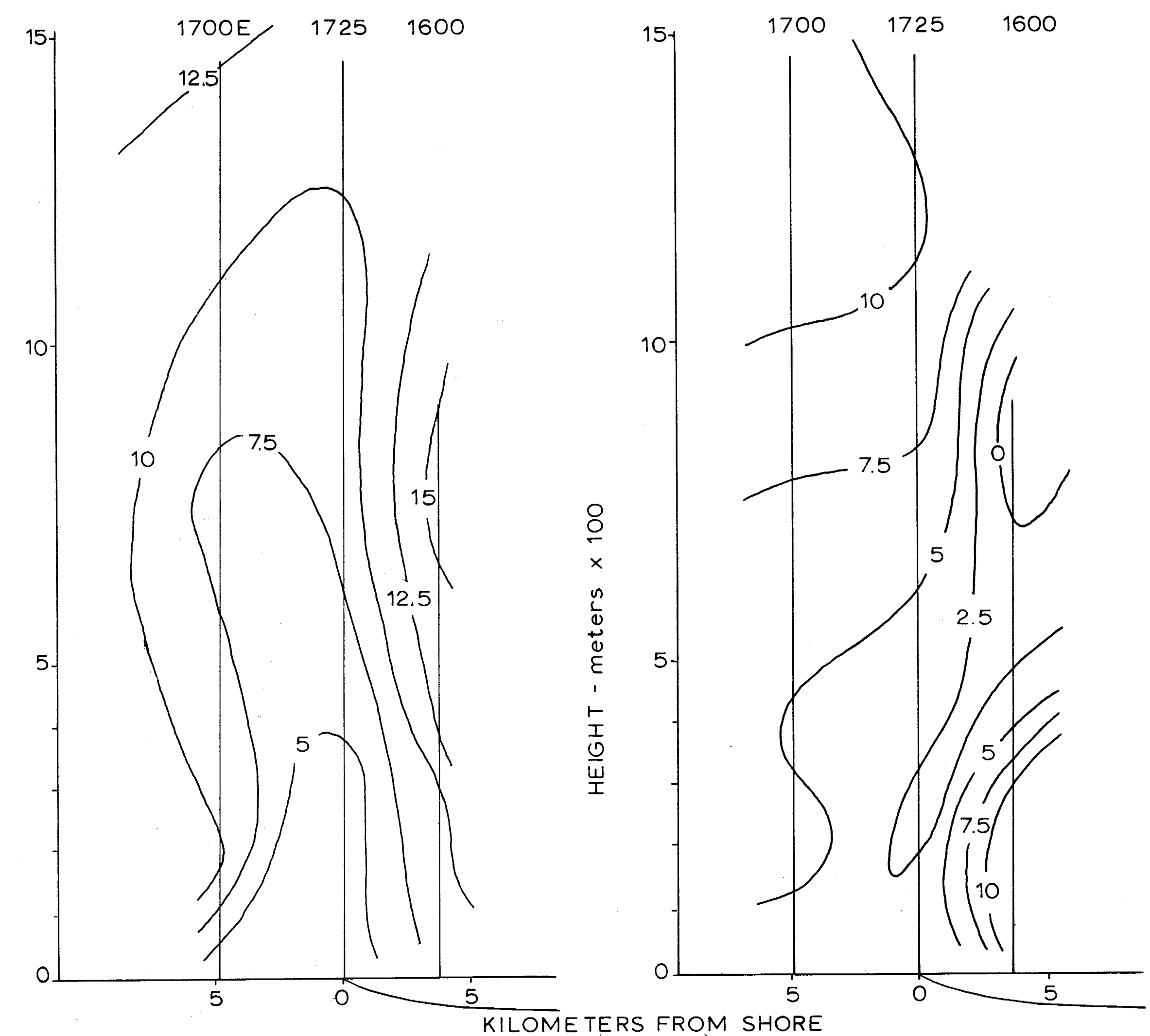

FIG. 6. U and V (eastward and northward) wind components. 8 June 1967, Waukegan, Illinois.

11 meters, verifies the strength of the inversion.

The effect of the lake's presence is seen in Figure 6 and appears to extend above 1500 meters. Subsidence over the lake at the upwind shore would designate the termination of any fair weather cumulus and possibly altocumulus (Lyons 1968), if present in the unstable atmosphere over land. A similar mechanism under certain conditions could suppress a squall line as mentioned by Lyons (1966).

The 1655E photo panoramic centered facing northward from Chicago, shown in Figure 5, shows the cumulus region and the sharp dissipation over the lake. Near the upper portion of the photograph several clouds are seen in the forming stages. As these move over the lake they are rapidly dissipated.

Regions of convergence or divergence in the horizontal wind field are representative of vertical motions in the atmosphere. Convergence close to earth's surface necessitates a vertical ascending current of air. Such convergence can be seen in Figure 4 between 5 kilometers inland and the shore. Eastward, between shore and 5 kilometers, extreme subsidence prevails below 1 kilometer as strong horizontal divergence is demanded from the difference between shore and MYSIS wind vectors.

Comparison of all our observations to Estoque's model (Figure 1 and 2) will demonstrate a high degree of agreement. Backing and veering winds across the shoreline appear in his model at close to observed distances from shore. Regions of divergence and convergence in the observations can be favorably compared with subsiding and ascending motions in the numerical model.

In all of the cases where actual observations are compared with Estoque's numerical results it becomes apparent that within the lower

wind levels agreement of numerical and observational results frequently fails. It is suggested that some degree of unrepresentativeness was introduced into the numerical model through the use of a constant flux layer in the inversion level (0-50 meters). The results indicate that a flux decay function down through the inversion might be more appropriate.

Alongshore gradient wind - high pressure over land

The northerly wind at Waukegan produces the second most dramatic case in the lake effect observations. One day typified this condition - 24 May 1967. Reference may be made to Figure A2 for the cloud picture and A11 for the 1300E meso-analysis. Generally there were scattered cirrus or cirrostratus throughout the day. The horizontal wind profiles have been separated into two periods: 1) 1015-1230E, Figure 7, and 2) 1340-1530E, Figure 8, to show the development of the lake effect under this wind regime. The lake breeze, which had already begun by period (1), became more intensified and better organized in the afternoon. The subjectively observed area of divergence and associated subsidence just offshore verifies the numerical results of Estoque as does the region of convergence immediately landward. Observations at the Four Mile Crib off Chicago verified the MYSIS observations of this near shore subsidence. (The Four Mile Crib is 4 statute miles east from the Chicago pierhead or 6.4 km.)

Since the observations were taken along a straight line it was considered undesirable to evaluate the lake effect for vertical motions. The similarity method used earlier (8 June-previous section) for approximating the w-component from two soundings was deemed unsatisfactory since several successive pibal measurement errors, due to vertical motions, would be magnified. The vertical motions resolved in this

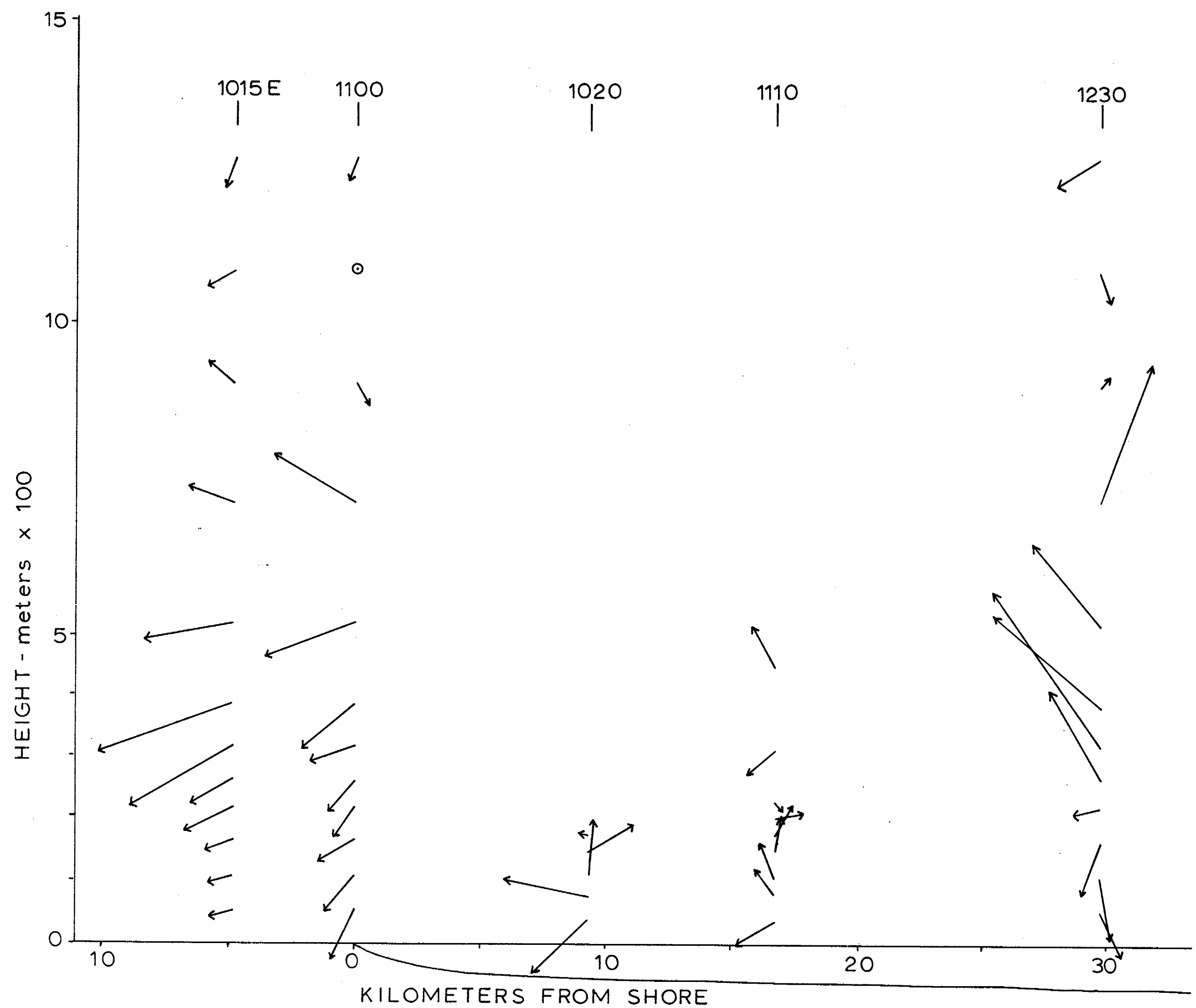

FIG. 7. Morning horizontal wind profiles. 24 May 1967. For scale see Fig. 4.

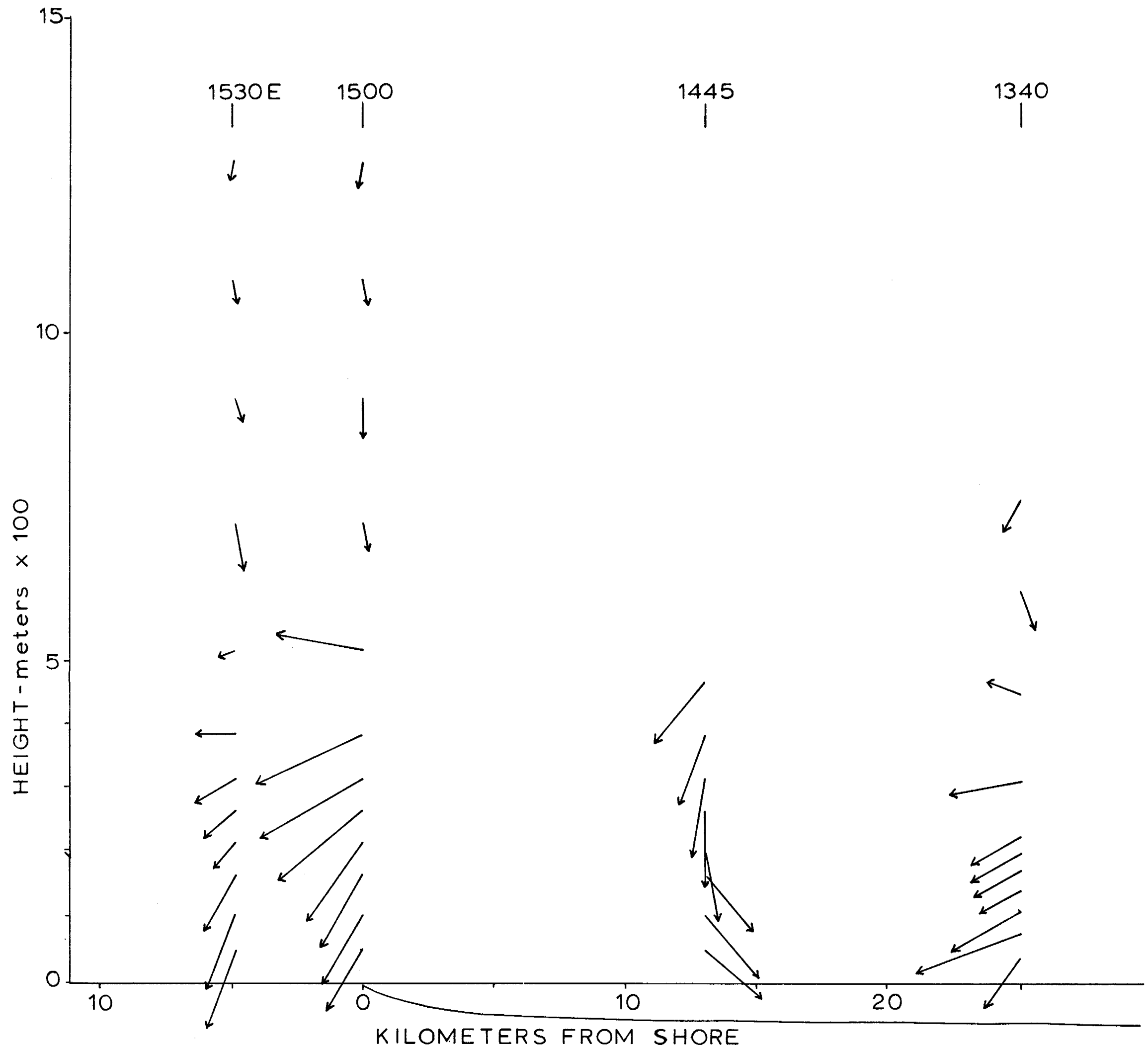

FIG. 8. Afternoon horizontal wind profiles. 24 May 1967. For scale see Fig. 4.

manner would be more meaningless than helpful. However, the first order estimations from subjectively interpreting convergent and divergent horizontal motions in the lower levels indicate positive and negative vertical velocities, respectively.

Wiresonde air temperature and dew point temperature measurements over the lake were made this day and are shown in Figures 9 and 10. They represent the mid-morning and mid-afternoon profile in the lowest 300 meters over the lake.[5] Since the soundings were not synoptic, the time scale has been added beneath the distance scale. The effect of the lake-induced higher pressure created a subsidence which reduced the inversion height by mid-afternoon. The top of the inversion was characterized by a change from cool moist air to a warming and drying with height. The anticyclonic strength of the lake atmosphere may be seen in the vorticity analysis at 1300E in Figure A22.

Under the parallel-to-shore gradient wind where higher pressure prevails landward, the diurnally induced lake high opposes the gradient wind. Subsequently, a northerly wind along the western shore of Lake Michigan will be balanced by the lake breeze to produce a narrow trough of lower pressure over land, close to and paralleling the shoreline. This lower pressure will be coupled with a ridge of higher pressure over the lake. The lower pressure along the western shoreline isolates the lake high from the continental high further west. Interaction of these high and low pressure features are illustrated in the 1300E meso-analyses for 26 May and 4 and 5 June (Figures A13, 17, 18). An identical situation should occur with a southerly gradient flow at the eastern shore.

[5] The maximum height of these soundings varied depending on the inversion characteristics.

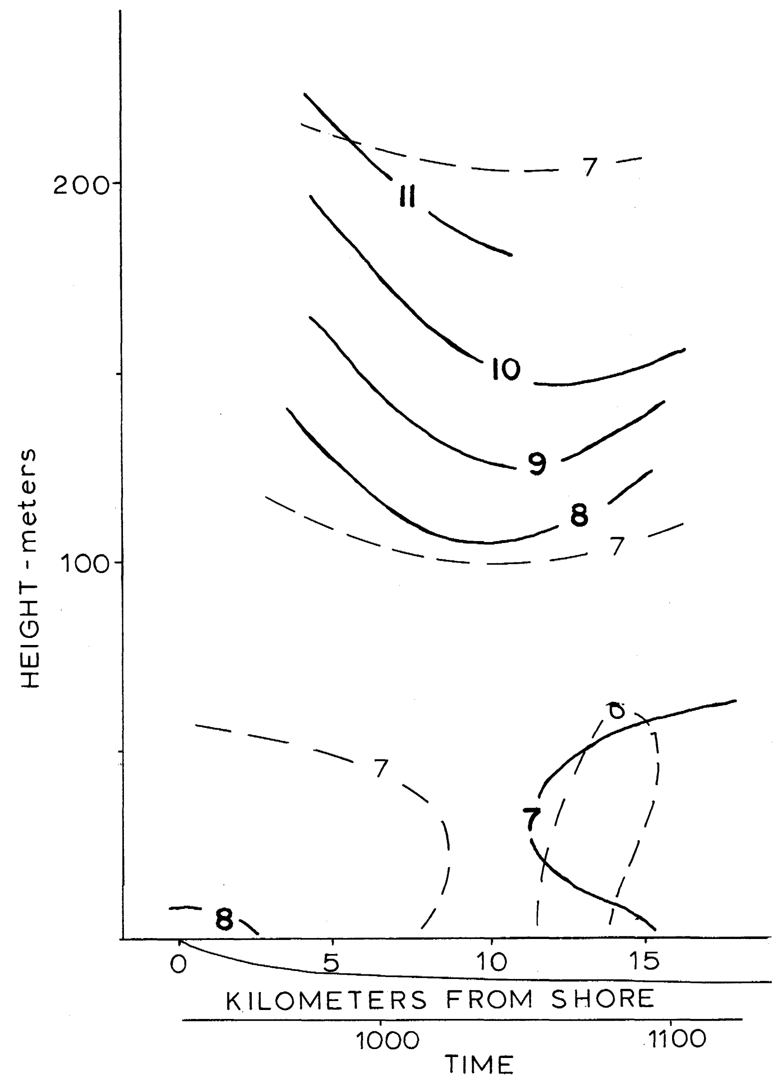

FIG. 9. Morning wiresonde profile of air (solid) and dew point (dotted) temperatures. 24 May 1967. East-west line from Waukegan, Illinois.

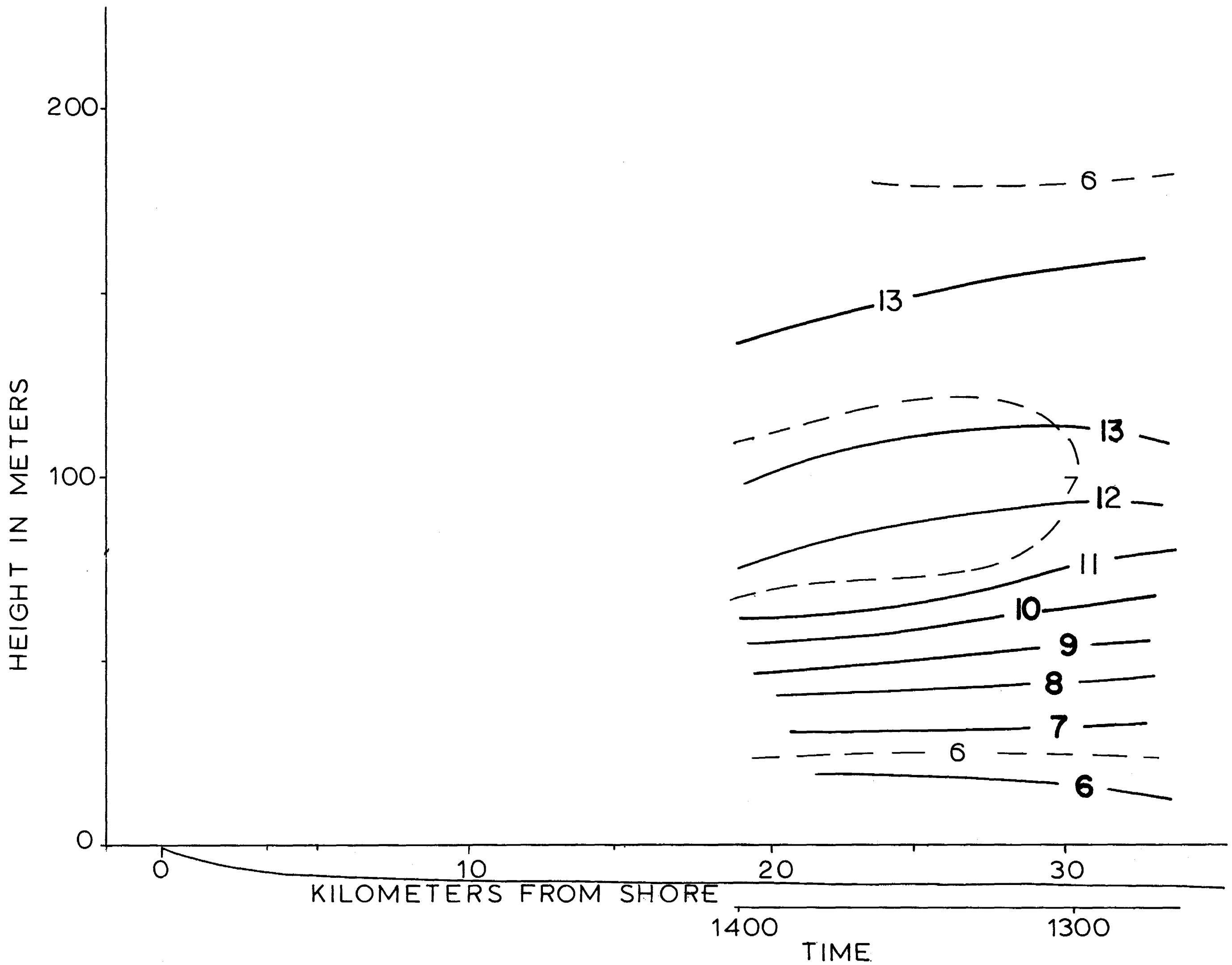

FIG. 10. Afternoon wiresonde profile. 24 May 1967. For details and conventions used see Fig. 9.

Onshore gradient wind

With an onshore gradient wind the vertical lake effect is minimal. Sensible heating from the land is carried inland by the lake air flow away from the land-water boundary. The lake air modification over land takes place slowly and results from a gradual increase in turbulence as the air moves over the warm soil. Three days of observations were made under this condition. 1, 2, and 3 June were quite similar in their onshore wind-flow characteristics. The gradient wind on 1 June was nearly twice those of the 2nd and 3rd - 12 m/sec, 5 m/sec, and 6 m/sec respectively. A high pressure cell to the north moved slowly south and eastward during these three days, keeping the skies nearly free of clouds - see Figure A5, A6, A7, A14, A15, and A16. The wind profiles, Figures 11 through 16, are for the morning and afternoon periods over the three days.

As observed in the previous case, a region of subsidence developed in the near shore region as the low level wind became oriented more parallel to the shoreline. Backing, down through the inversion, of the horizontal wind vector was restricted below 500 meters in the morning but expanded by mid-afternoon to affect the flow to 1000 meters.

Effect of the stronger gradient on 1 June was reflected in a reduced vertical extent of the lake effect. In all three observations of the onshore wind a 90° backing in direction of the low level wind was produced down through the inversion over the lake. This feature appeared to be most pronounced in a region 20-30 kilometers from shore.

True to the form of all the remaining days of the study, the wind profile demonstrated a low level peak within the first 300 meters (quite typically 150 meters). Above this level, a decrease in wind speed was customarily followed by an increase toward the upper limit of the pibal

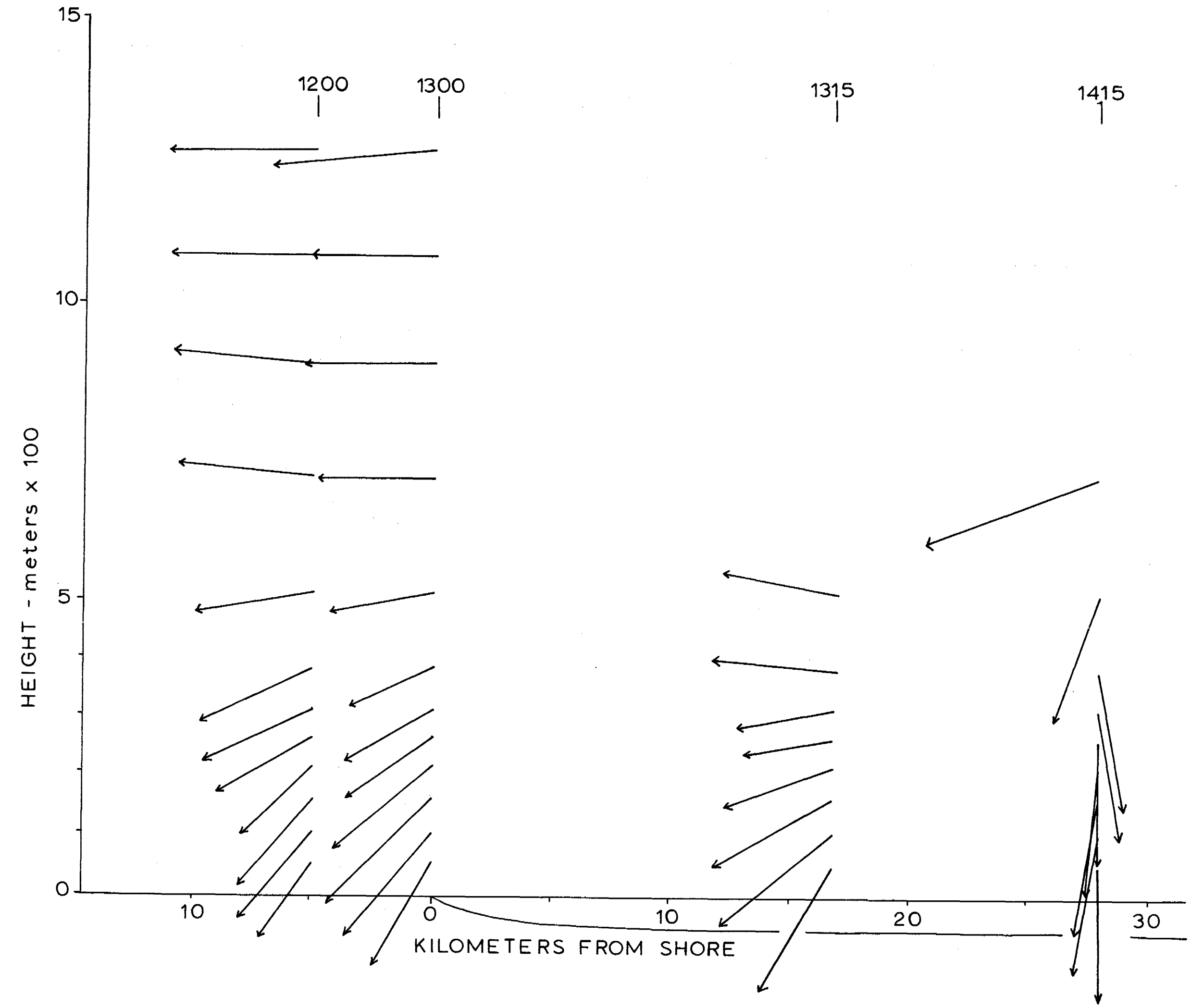

FIG. 11. Early afternoon horizontal wind profiles. 1 June 1967. For scale see Fig. 4.

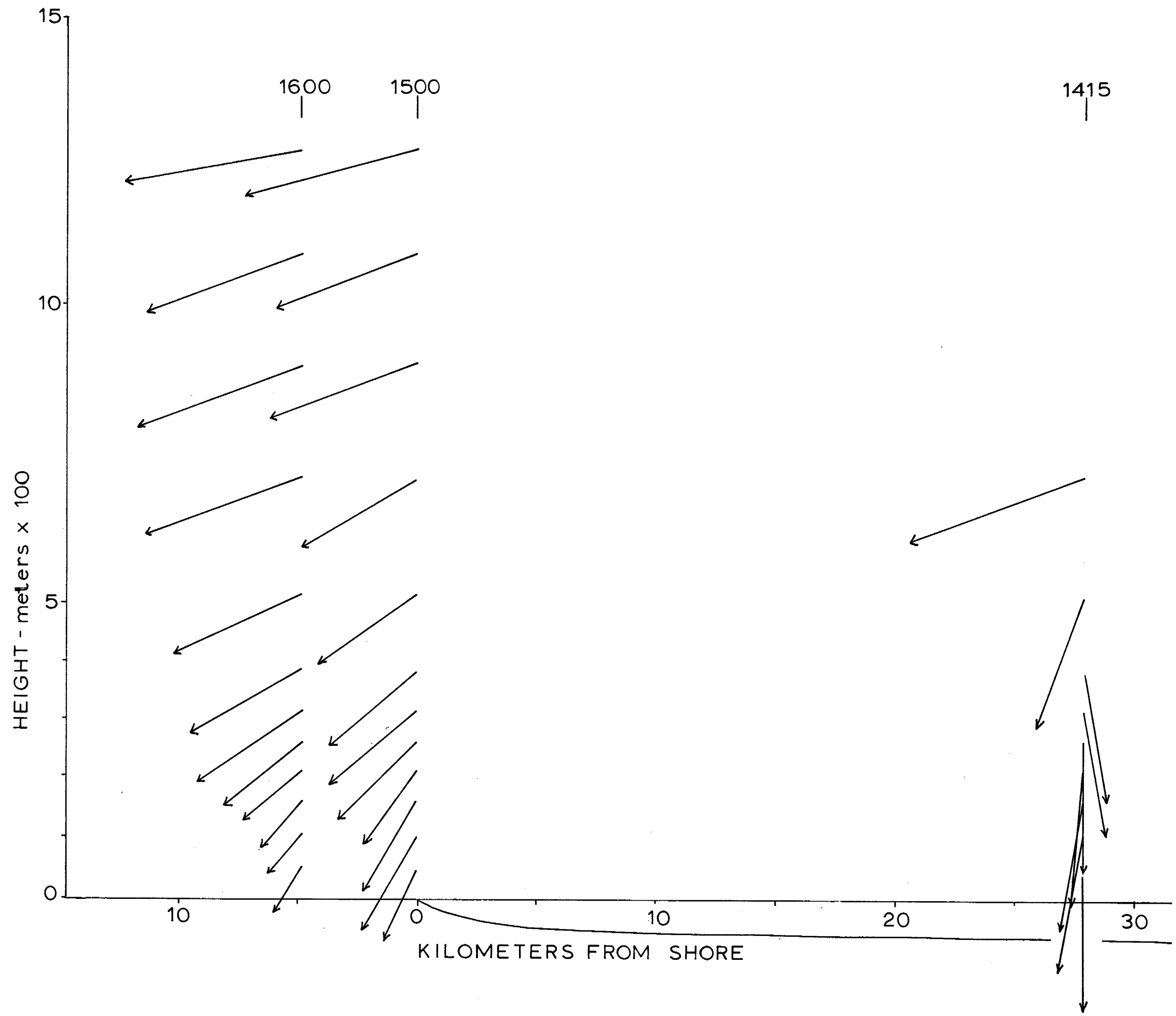

FIG. 12. Late afternoon horizontal wind profiles. 1 June 1967. For scale see Fig. 4.

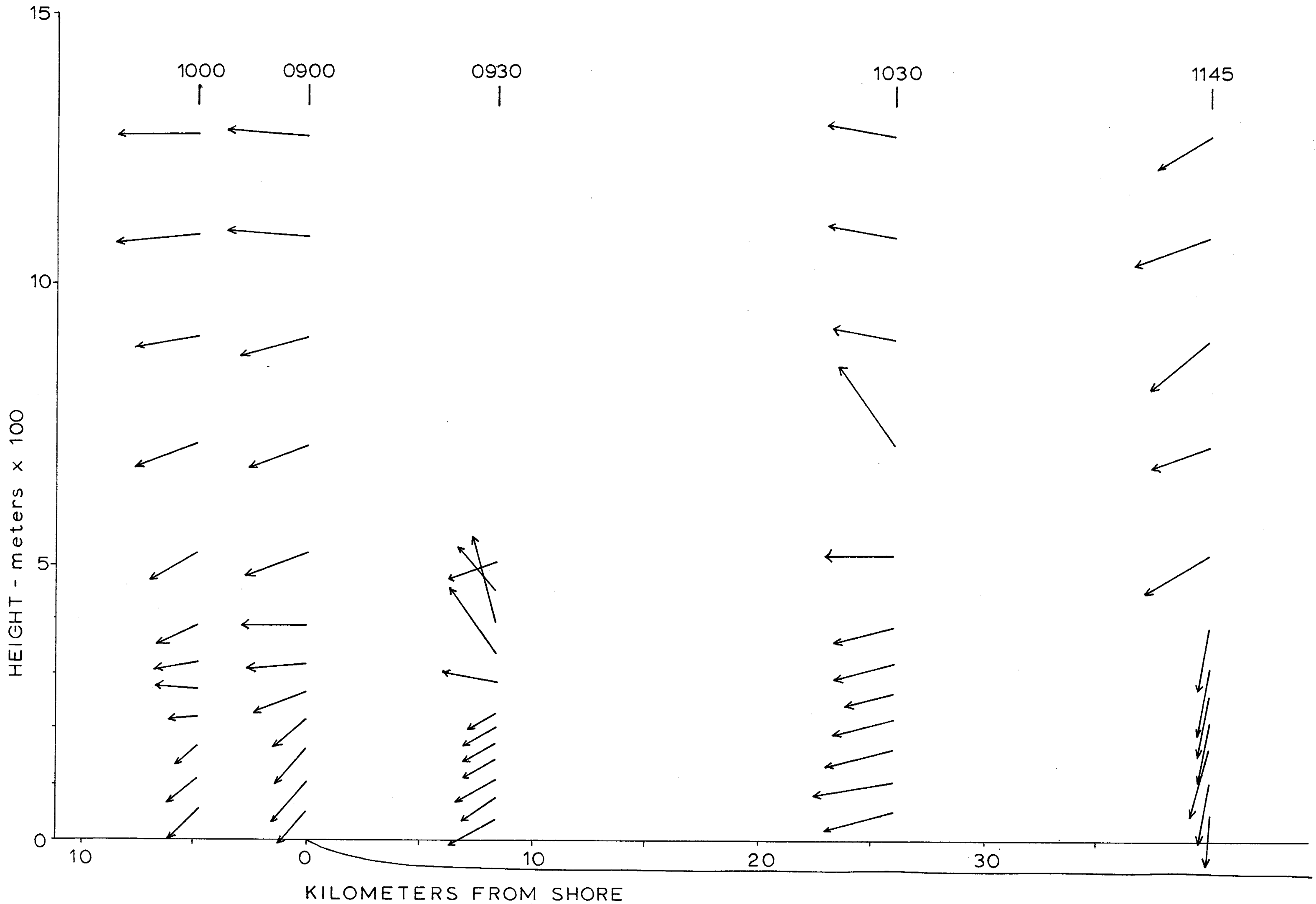

FIG. 13. Morning horizontal wind profiles. 2 June 1967. For scale see Fig. 4.

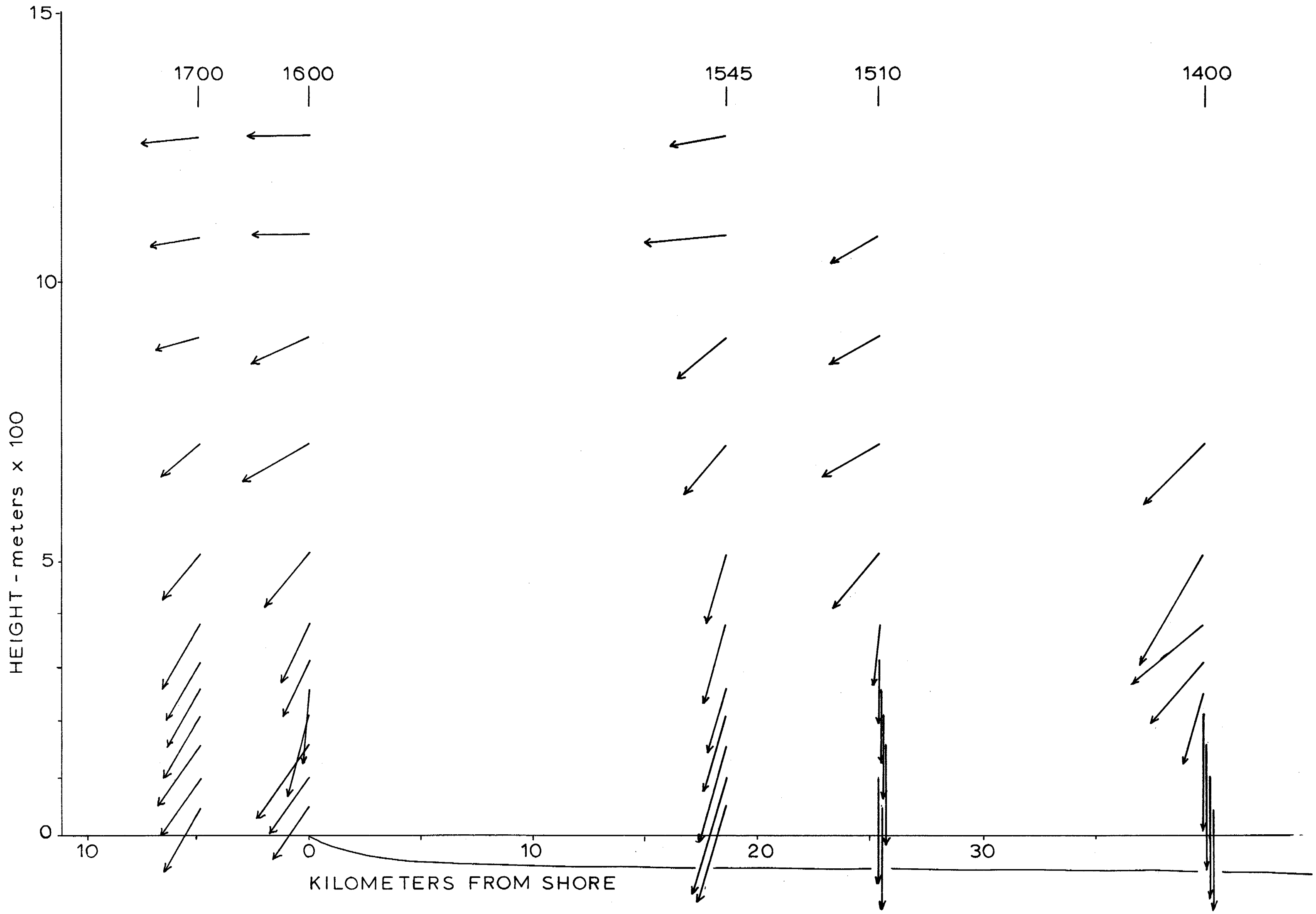

FIG. 14. Afternoon horizontal wind profiles. 2 June 1967. For scale see Fig. 4.

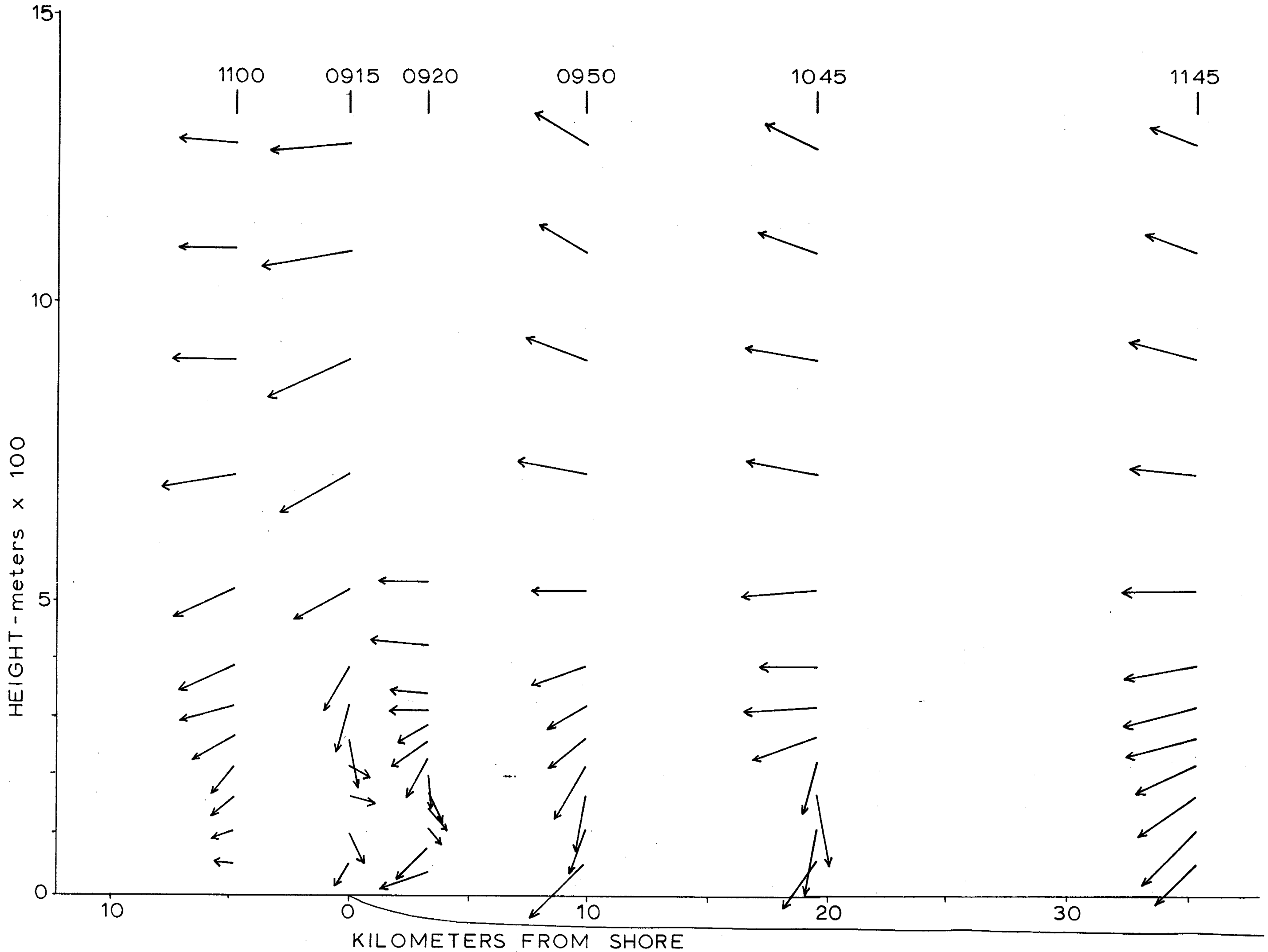

FIG. 15. Morning horizontal wind profiles. 3 June 1967. For scale see Fig. 4.

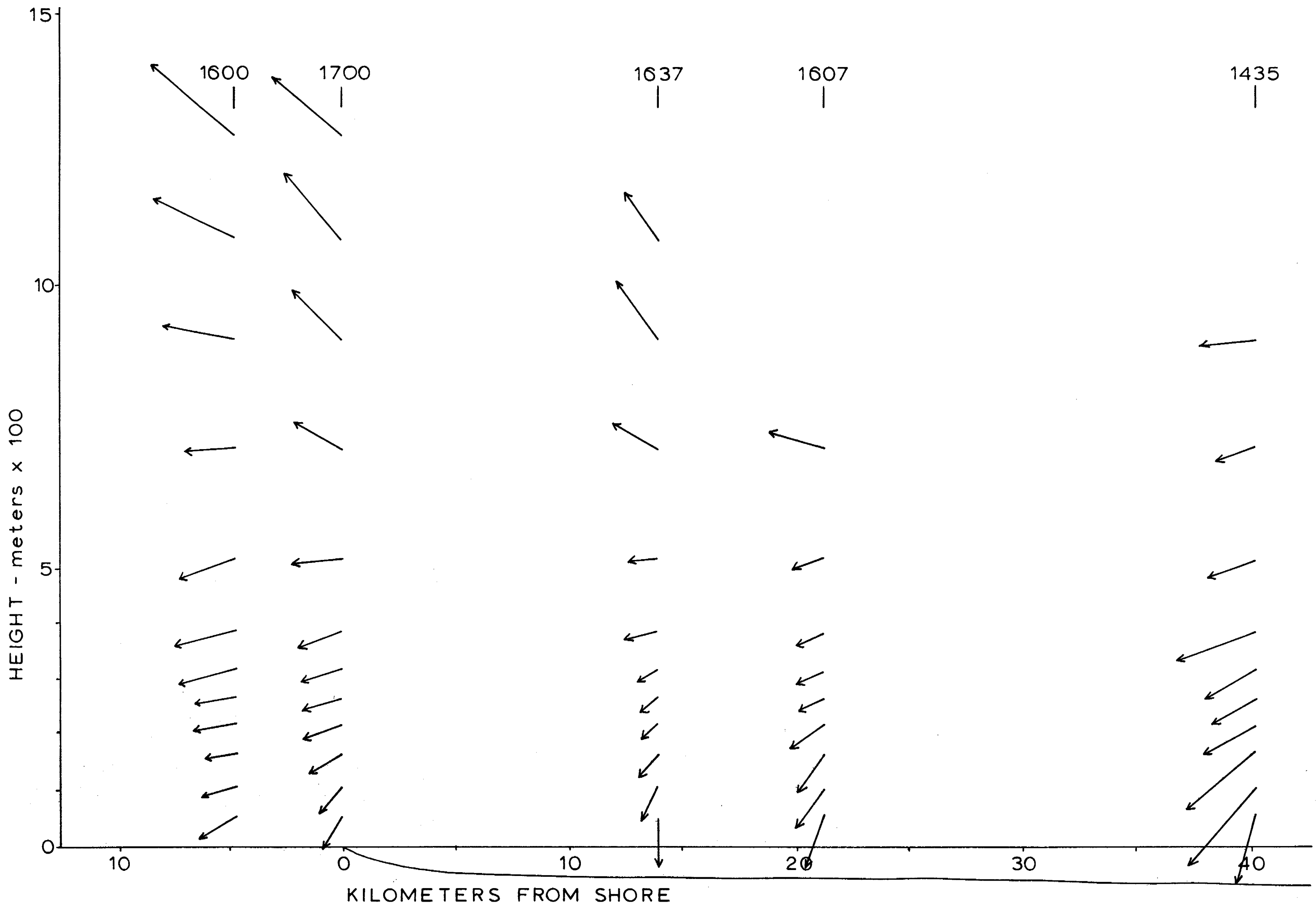

FIG. 16. Afternoon horizontal wind profiles. 3 June 1967. For scale see Fig. 4.

sounding (1500 meters). These distinct features are representative of the lake effect (150 meter peak) and the gradient flow aloft.

On both the 2nd and 3rd, wiresonde measurements of the onshore wind condition were obtained. These cross sections are presented in Figures 17 and 18 for 2 June and Figures 19 and 20 for 3 June in the same manner of presentation as earlier figures (9 and 10). Both days evinced inversion intensification as a result of subsidence in the meso-scale circulation of the lake anticyclone. The differences on the 3rd were most probably a reflection of the increased southerly component of the onshore gradient wind. This day's morning dew point profile appeared very similar to a vapor pressure cross section published by Craig, Katz, and Harney (1945). In their study made over Massachusetts Bay, the vapor pressure in the lowest 300 meters showed the lifting of low level moist sea air as it moved over land, and the subsidence over the cool Bay of warmer and drier air aloft. At Waukegan, as the lake effect intensified during the afternoon, this effect became obscured.

Comparison of this wind condition to the Estoque model can be visualized at the righthand shoreline of Figure 1. The X axis and normal Y axis orientation needs to be reversed in sign to provide the onshore flow lake effect at the western shore of Lake Michigan. A wind vector directed into the model (positive v component) would point southward in a reversed image of Figure 1 and the gradient flow would become an easterly one. The onshore flow shows much less effect in the vertical than at the across-lake or upwind shore. Estoque shows some backing of the wind within the inversion. Observations show this backing is much greater.

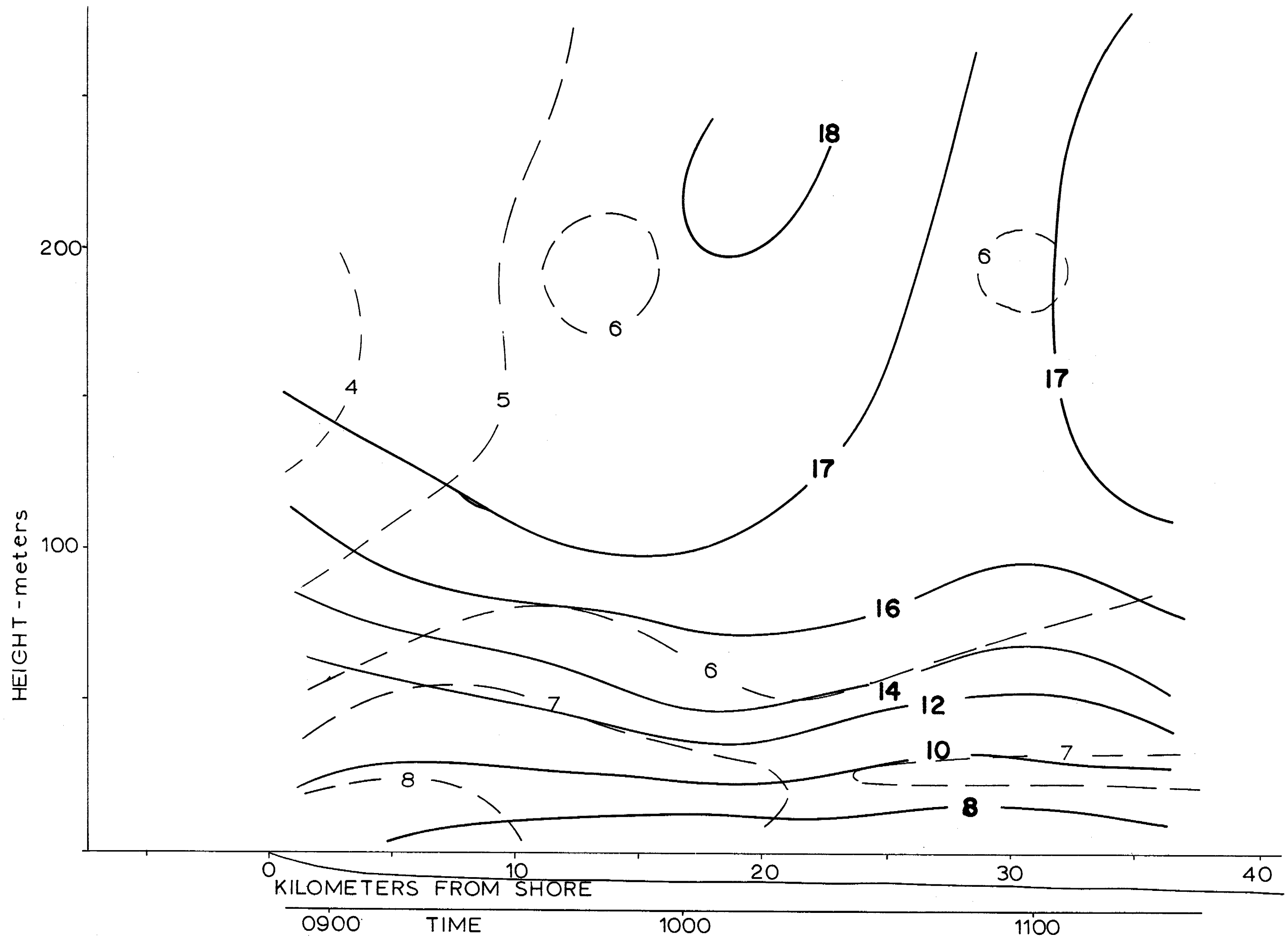

FIG. 17. Morning wiresonde profile. 2 June 1967. For details and conventions used see Fig. 9.

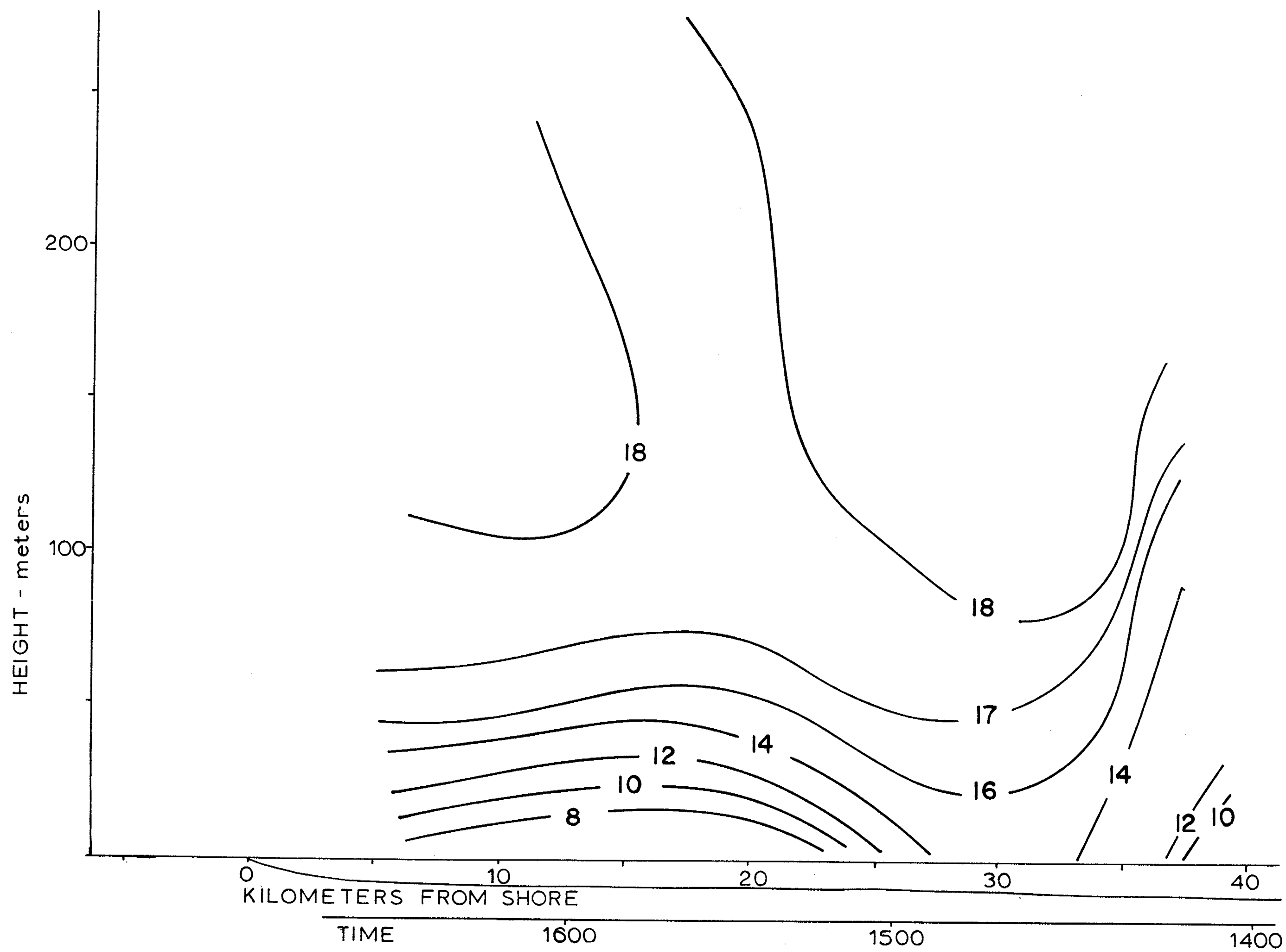

FIG. 18. Afternoon wiresonde profile. 2 June 1967. For details and conventions used see Fig. 9. [No dew point temperatures available.]

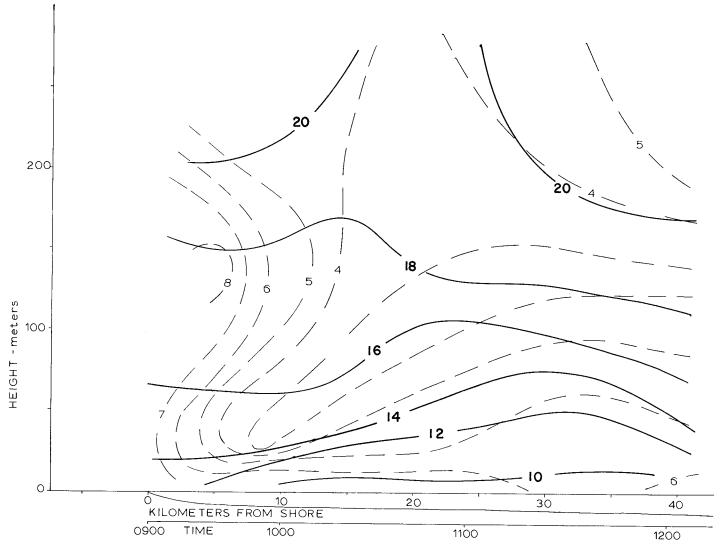

FIG. 19. Morning wiresonde profile. 3 June 1967. For details and conventions used see Fig. 9.

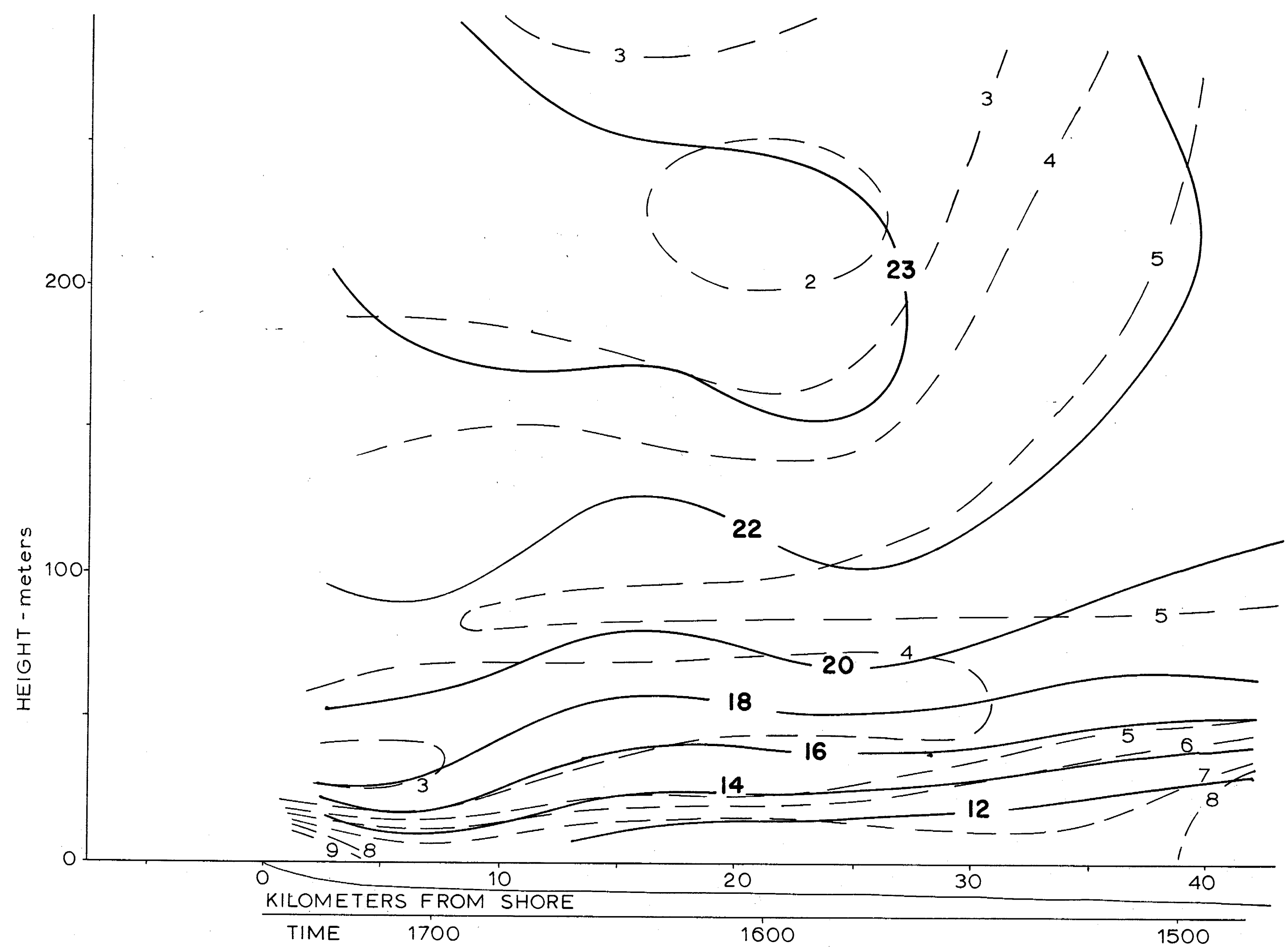

FIG. 20. Afternoon wiresonde profile. 3 June 1967. For details and conventions used see Fig. 9.

Munn (1966) has discussed the boundary layer condition which results as the lake air is heated from below once it begins moving over the land at the downwind shore. The lapse rate of the lake air in this region becomes progressively more superadiabatic and, depending largely as a function of the strength of the induced convection and the gradient wind speed, gradually erases the inversion condition. Once the inversion is dissipated, turbulent processes convert the flow to a more truly characteristic land flow of forced convection. Cumulus development will no longer be suppressed and these clouds commonly form in cloud streets along a locus parallel to the shore (Lyons 1966).

Alongshore gradient wind - high pressure over lake

The three days characterized by the gradient wind parallel to the coastline, with higher pressure lakeward, have been summarized in Table 3: 25 May, 4 and 5 June. The strongest gradient wind - 8 m/sec - occurred on 4 June. The lake effects on this particular day may have received an additional influence due to the proximity of a warm front approaching from the west. However, the presence of the lake effect, and particularly of the lake high (see Figure A17), appeared instrumental in slowing the advance of the front during the day.

In the same manner as in the preceding section, the horizontal wind vectors are presented in Figures 21-26. In all the observations of this condition the onshore flow in the lower levels increased during the morning in both intensity and in vertical extent. The wind velocity in the vertical wind profile peaked around 300 meters and the direction gradually veered around to more gradient conditions at a height of 1200 meters above the surface. Alternating convergent and divergent motions are particularly evident in Figure 24 (4 June). These tendencies are believed to be coupled

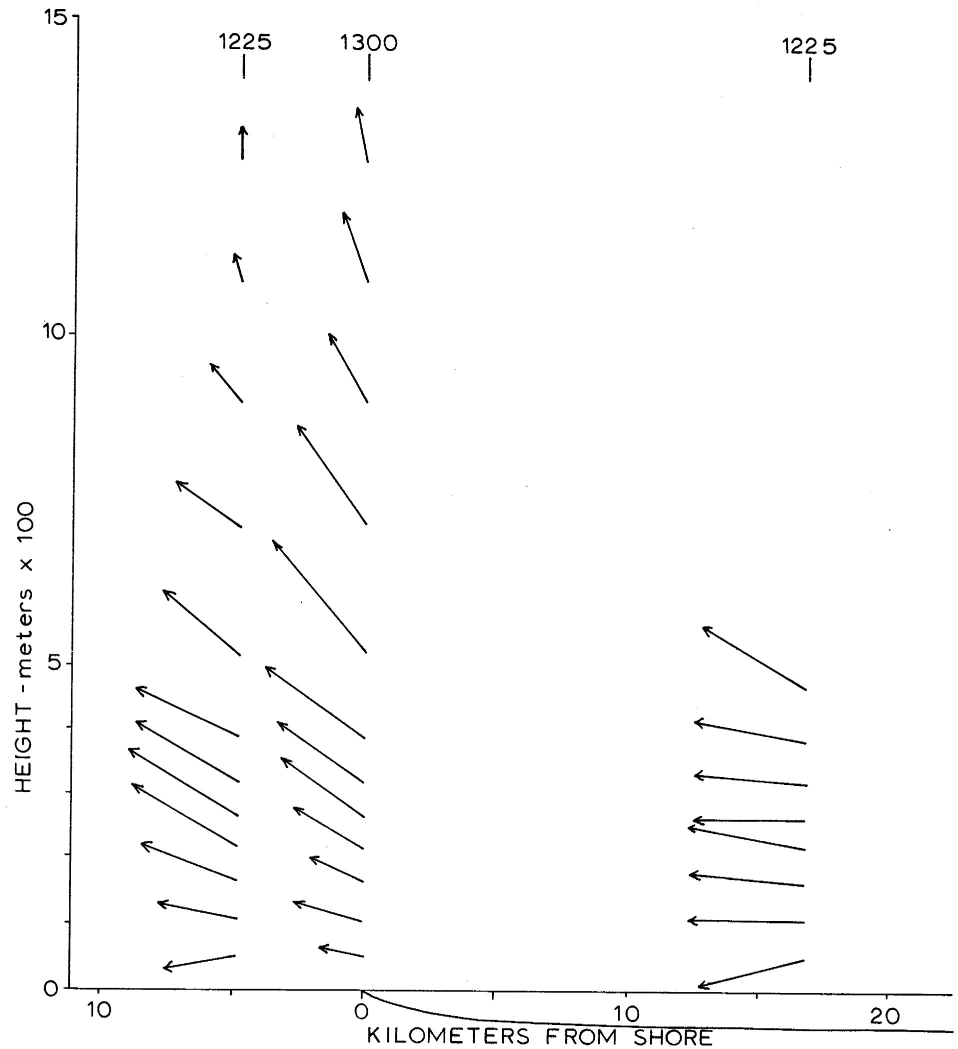

FIG. 21. Early afternoon horizontal wind profiles. 25 May 1967. For scale see Fig. 4.

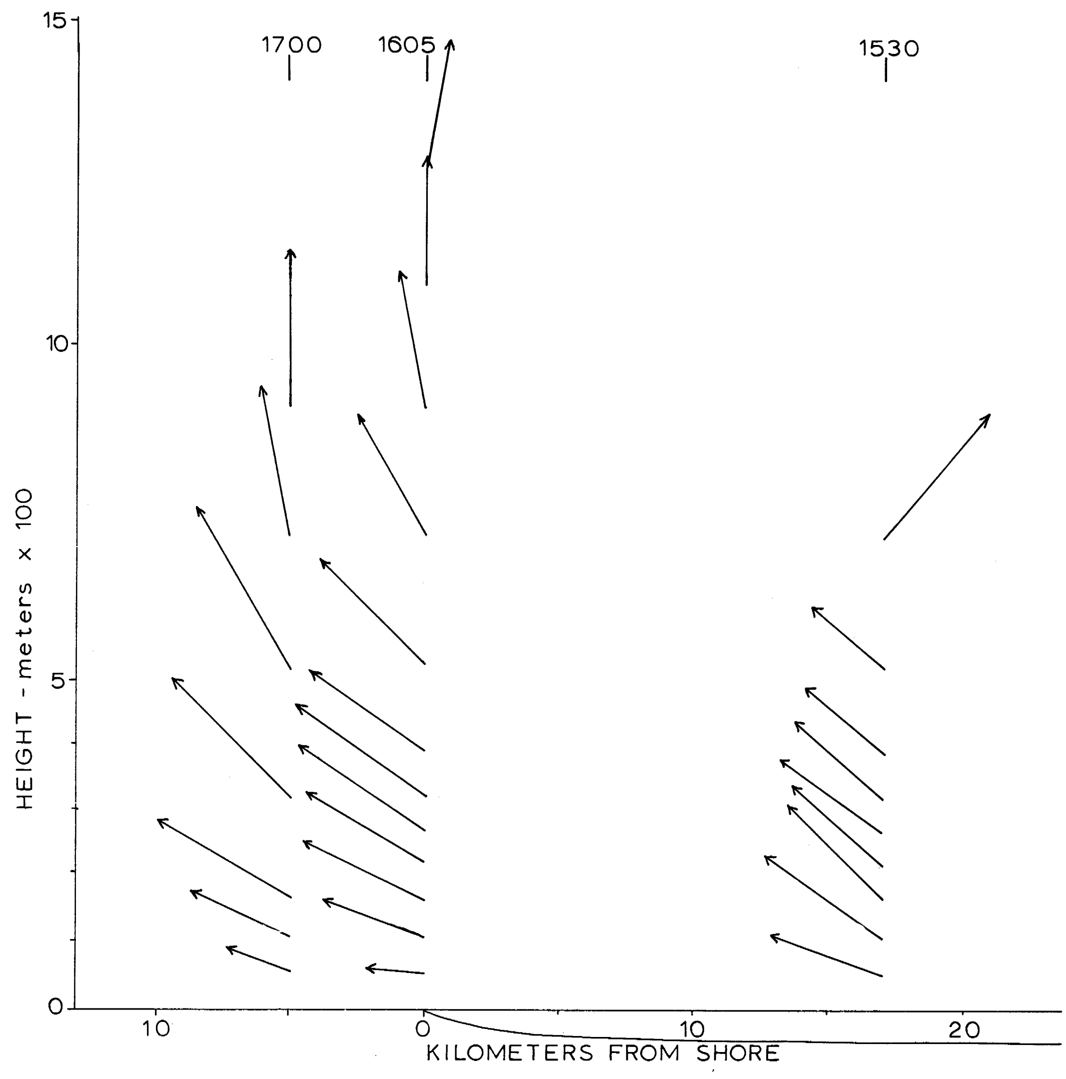

FIG. 22. Late afternoon horizontal wind profiles. 25 May 1967. For scale see Fig. 4.

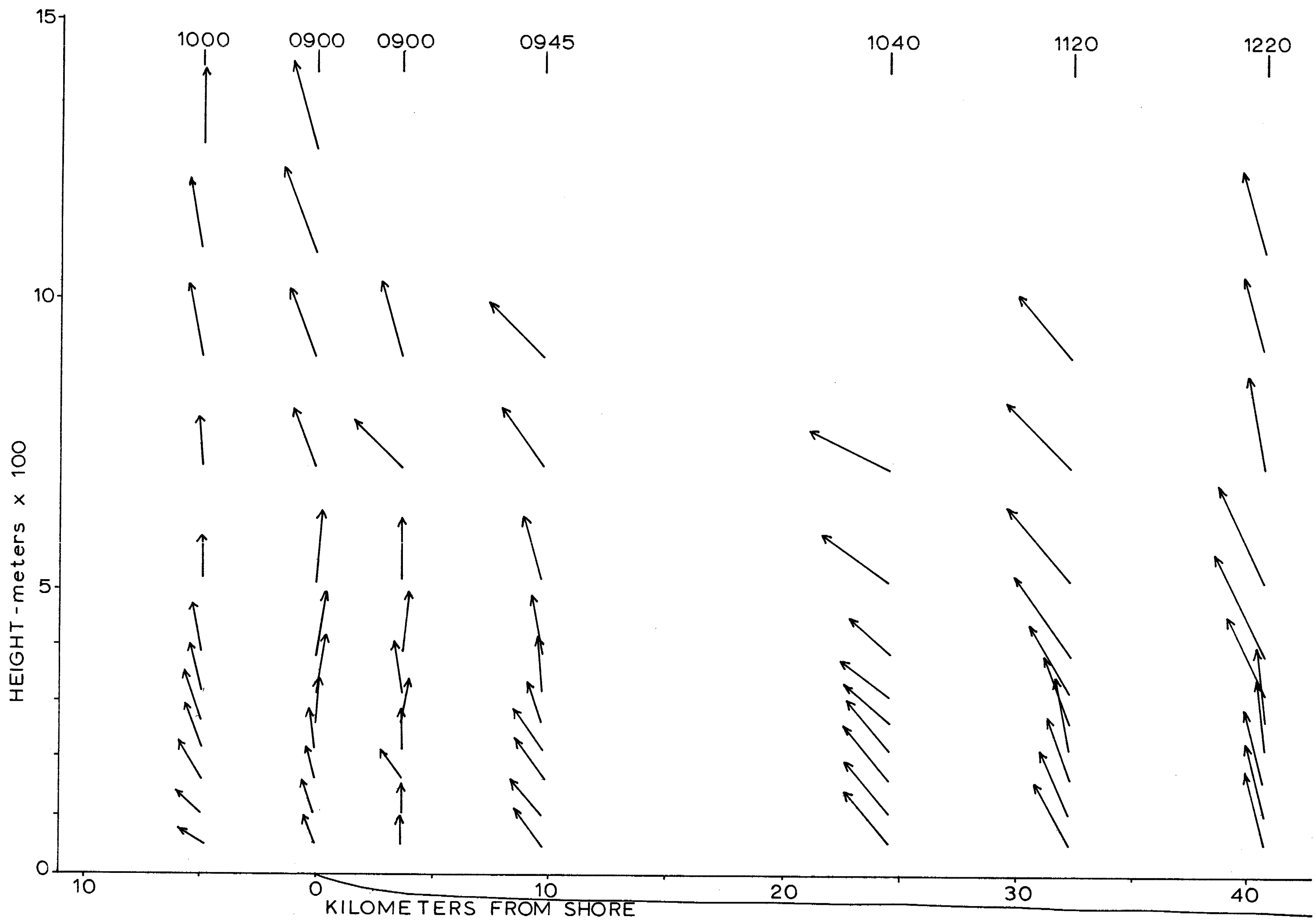

FIG. 23. Morning horizontal wind profiles. 4 June 1967. For scale see Fig. 4.

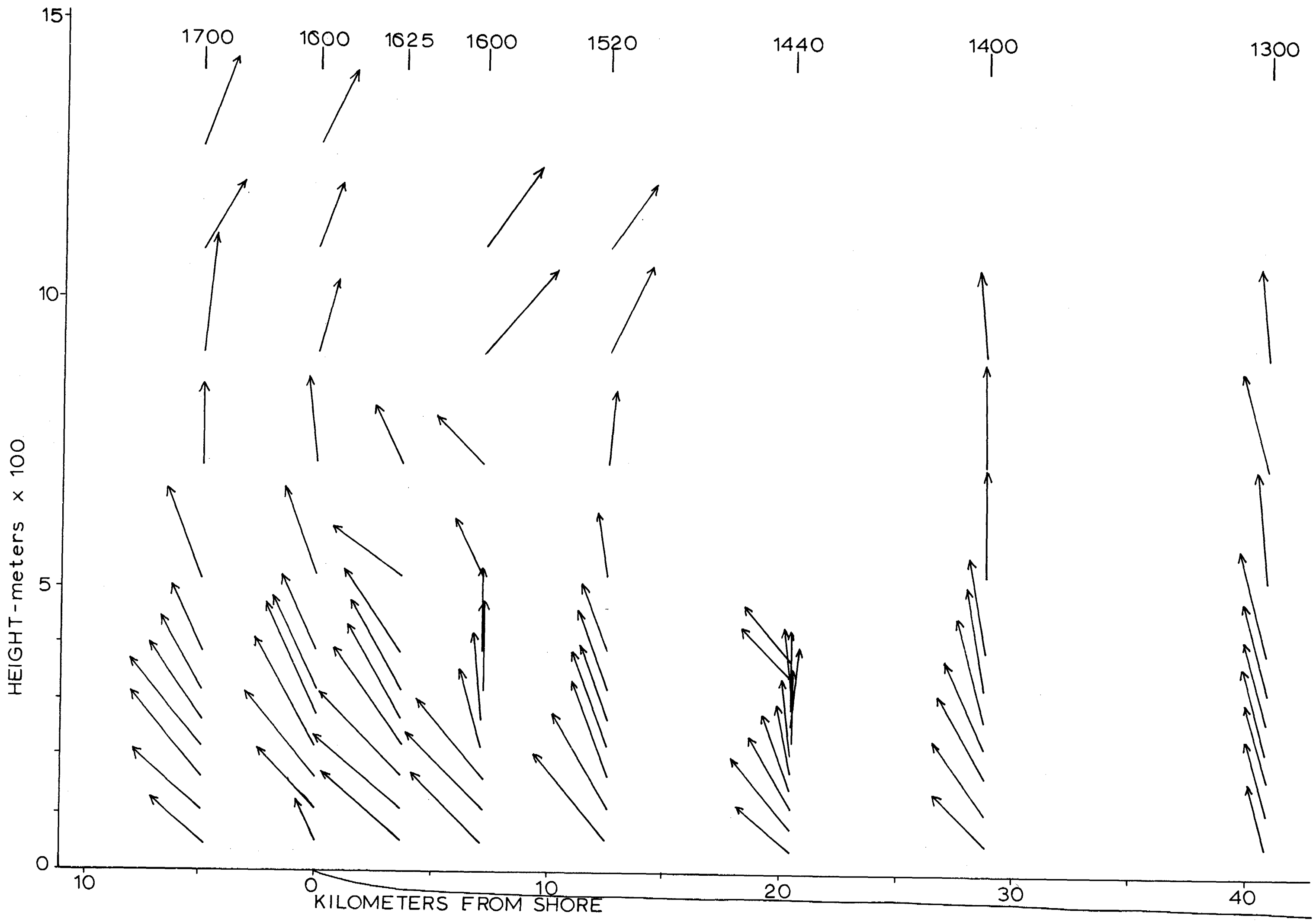

FIG. 24. Afternoon horizontal wind profiles. 4 June 1967. For scale see Fig. 4.

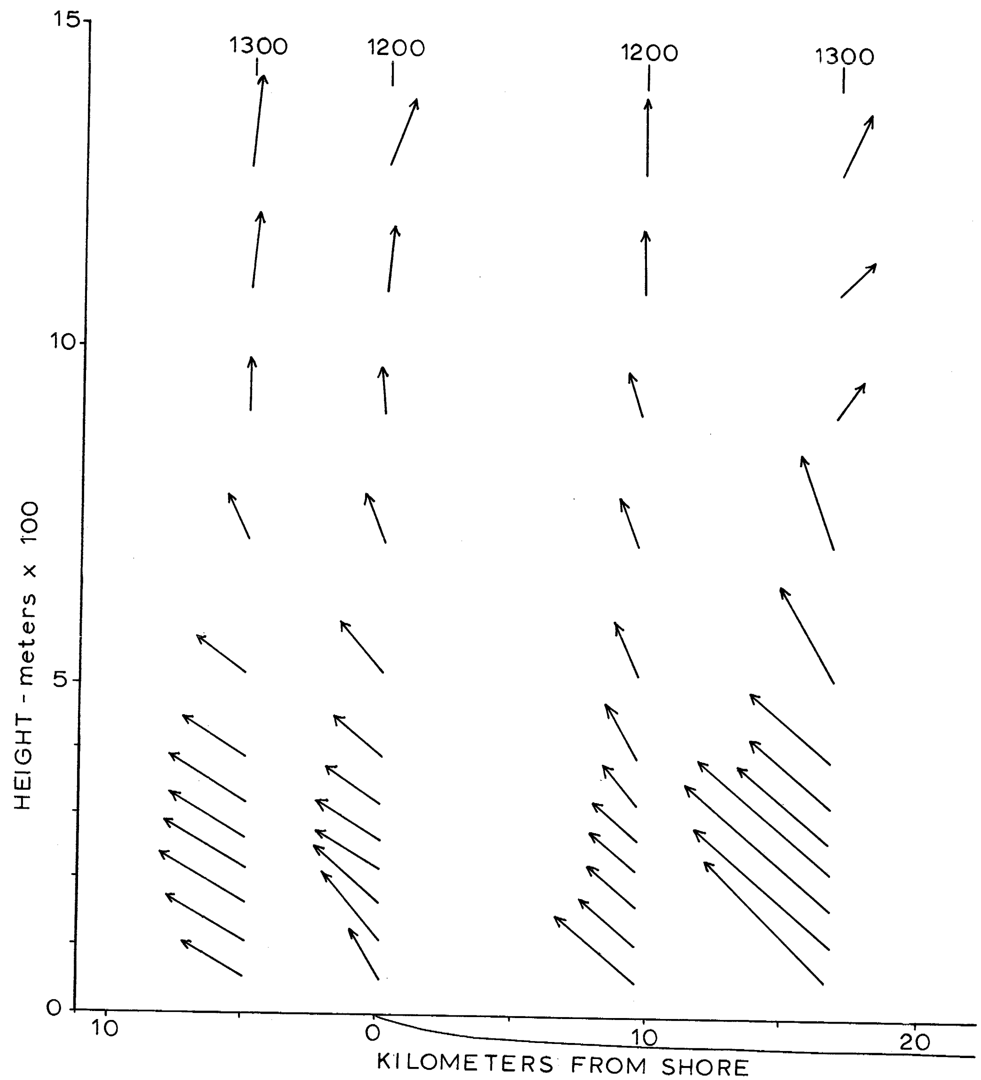

FIG. 25. Early afternoon horizontal wind profiles. 5 June 1967. For scale see Fig. 4.

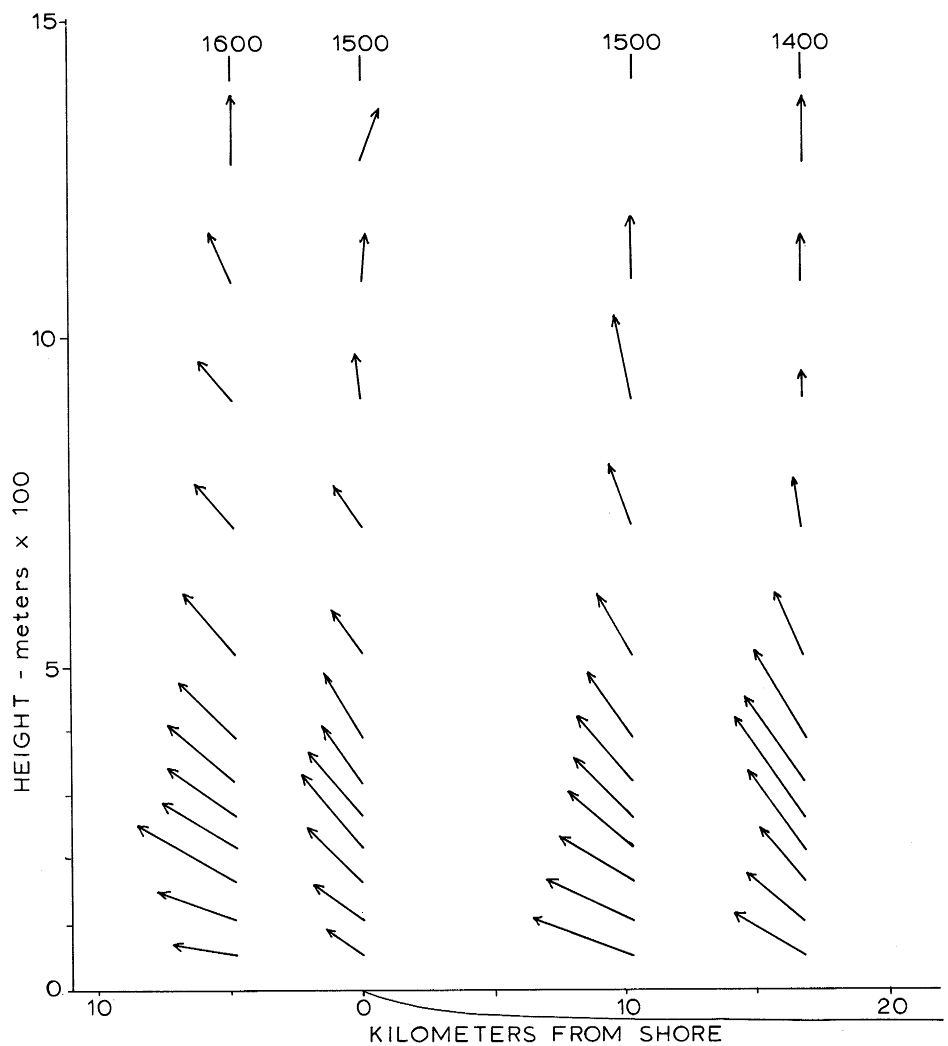

FIG. 26. Late afternoon horizontal wind profiles. 5 June 1967. For scale see Fig. 4.

with internal waves initiating at the top of the inversion, which was close to 100 meters. This phenomenon has previously been reported in sea breeze measurements by Donn et al. (1956).

The air and dew point cross sections show inversion features. Figures 27 through 30 present these data of late morning and mid-afternoon measurements over the three days. As seen previously, wide-spread subsidence intensified the thermal gradient during each day as the parallel-to-shore gradient wind became altered to a landward flow at the surface and lakeward flow above 1200 meters. It can be seen in a comparison to the opposite parallel-to-shore gradient wind that less vertical motion occurred near shore under this final condition. Such a result should be expected as the flow direction (southerly) implies higher pressure lakeward which is in accord with the lake induced anticyclone.

The modified Estoque lake effect model has already provided a vertical cross section of the lake under prevailing parallel-to-shore gradient wind (Figure 2). More uniformity in flow is characteristic at the shore where the gradient wind implies higher pressure lakeward (e.g. northerly gradient wind over the eastern shore or southerly gradient wind over the western shore).

Summary of lake effect winds

The warm offshore gradient wind is most dramatically affected by the cold lake. While the horizontal extent of these effects is limited there is strong indication that the vertical extent is quite large. A region of rapid upward motion over land is followed by a strong subsidence near shore and over the water. While this characteristic represents a wave in the atmosphere over the shoreline it will become a lake breeze

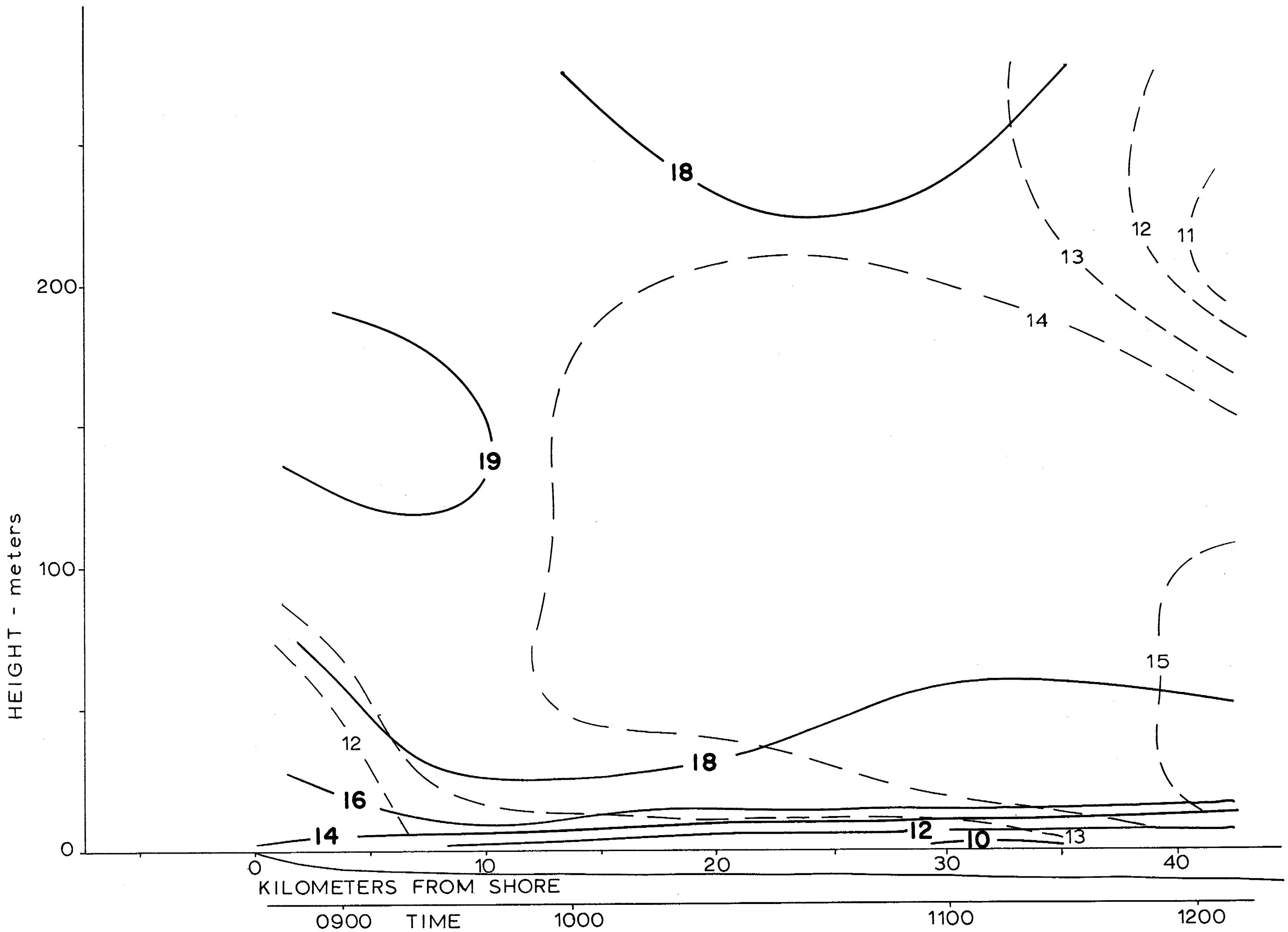

FIG. 27. Morning wiresonde profile. 4 June 1967. For details and conventions used see Fig. 9.

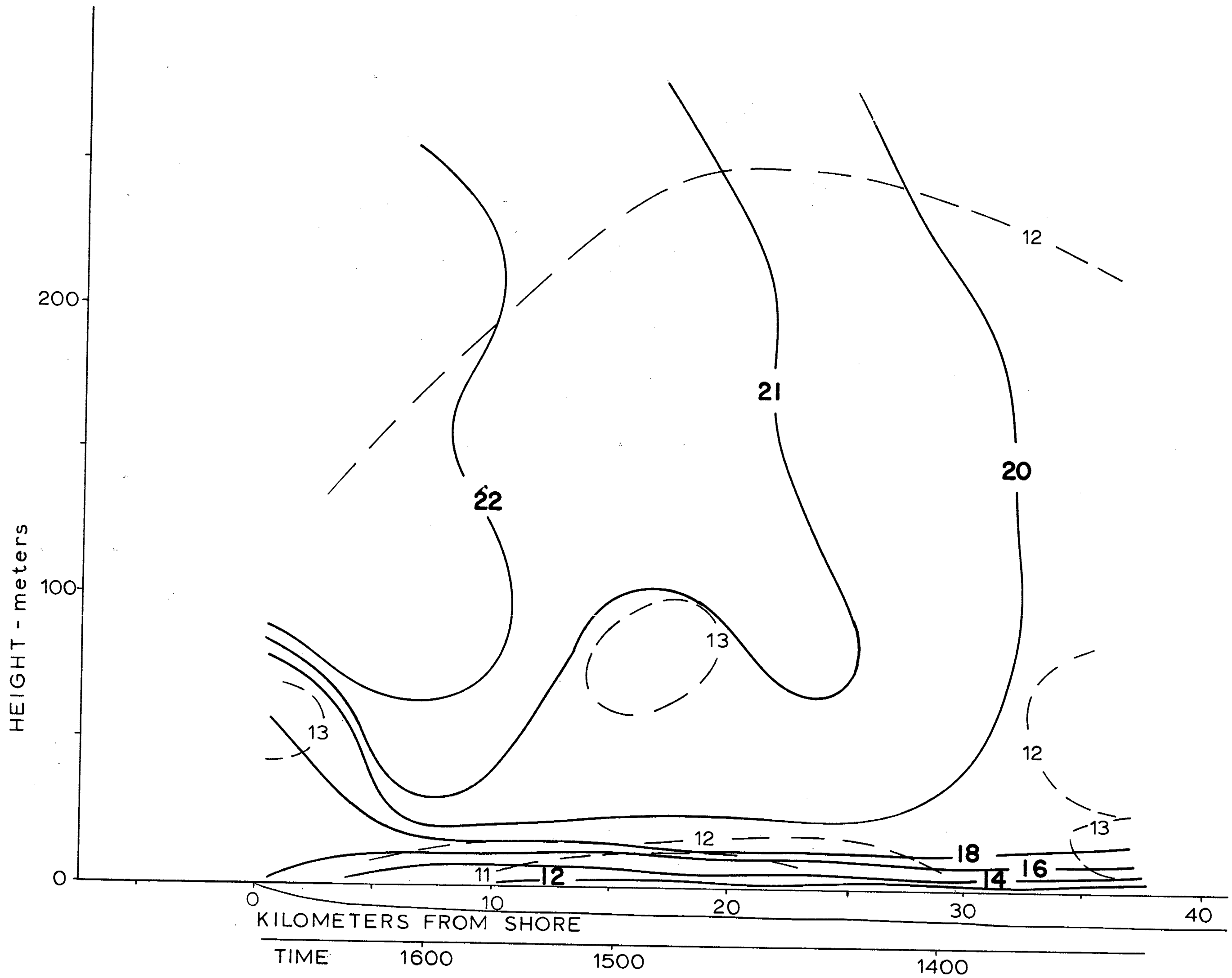

FIG. 28. Afternoon wiresonde profile. 4 June 1967. For details and conventions used see Fig. 9.

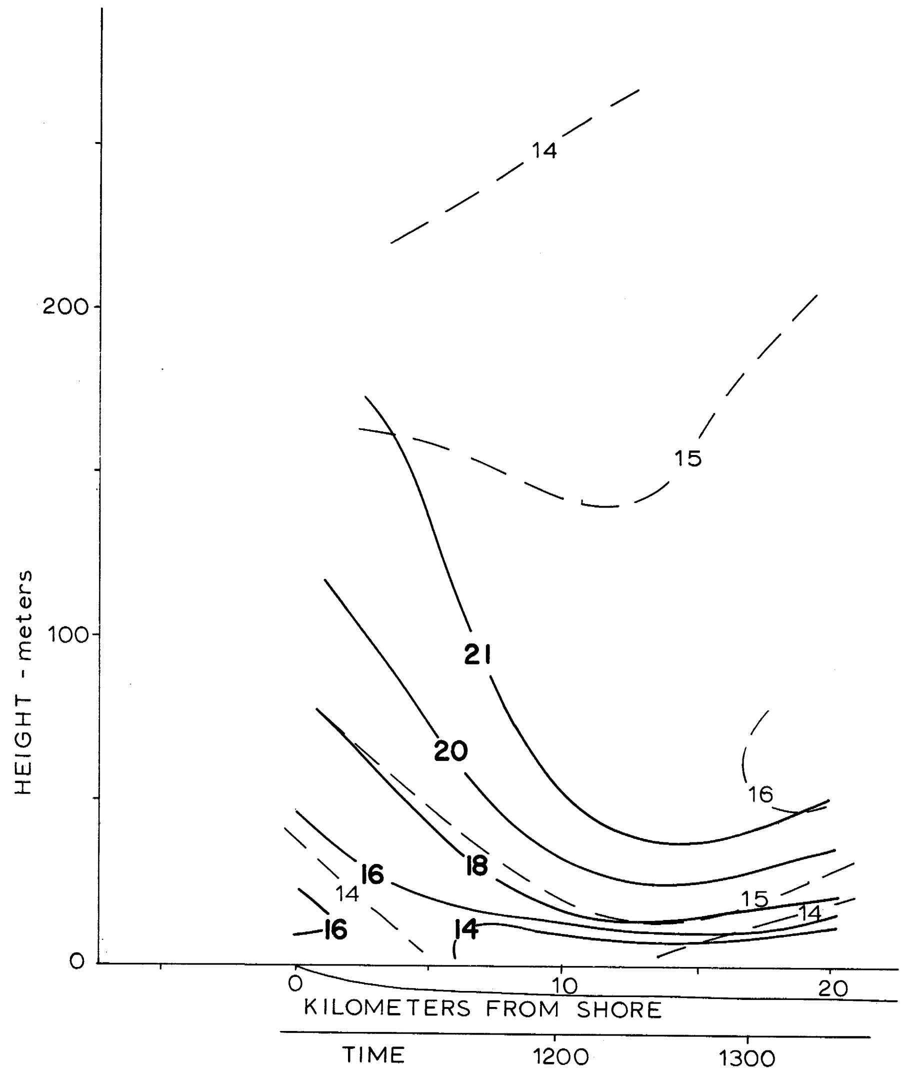

FIG. 29. Early afternoon wiresonde profile. 5 June 1967. For details and conventions used see Fig. 9.

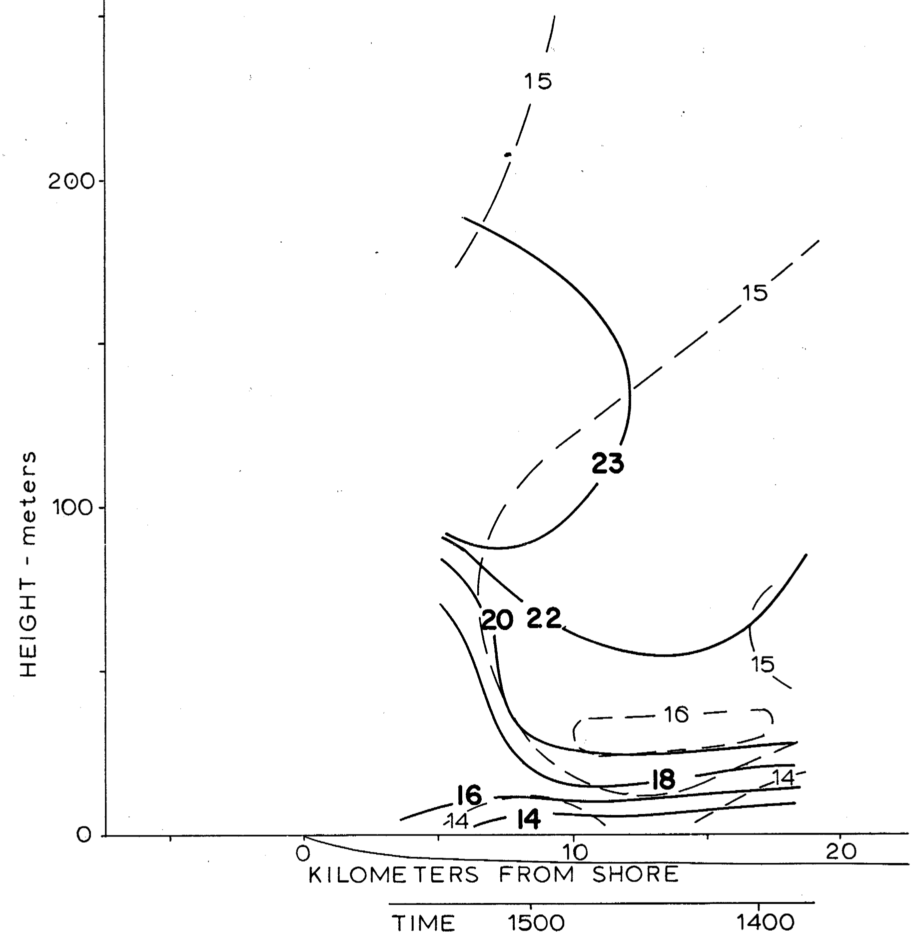

FIG. 30. Late afternoon wiresonde profile. 5 June 1967. For details and conventions used see Fig. 9.

circulation if the pressure gradient is weak. The wave forms the leading edge of the inversion over the lake and may be important in setting up strong internal waves in the inversion. Interesting work lies ahead in connection with this offshore wind condition. Double theodolite wind soundings taken simultaneously from shore and some fixed object several kilometers over the water (e.g. a water intake crib) should produce valuable information on this extreme lake effect phenomenon.

In contrast to the offshore gradient wind, the onshore gradient wind is least affected by the lake as most of the energy is being dissipated at the opposite (upwind) shore. The inversion is better developed, allowing the wind to veer to a direction that may be more nearly parallel to the downwind shore. Vertical velocities at the downwind shore are small. As the inversion is carried landward the lake air modifies as the air heats over the warmer soil and returns the vertical temperature profile to a neutral lapse rate. By the time the air has been warmed to an unstable lapse rate, cumulus may develop marking the horizontal limit of the lake effect.

Parallel-to-shore gradient wind conditions produce complex lake effect winds when the driving pressure gradient wind results from lower pressure lakeward and higher pressure landward. As the lake associated high builds, during the day, the higher pressure lakeward causes a region of low pressure to exist over the much warmer shoreline. This low pressure trough partially separates the gradient wind high pressure system from that of the lake high. The combination of high and low pressure cells permits the balancing interaction of the lake effect to be carried to higher levels. Cloud patterns may be affected along and near the shoreline as low level convergent and divergent cells operate in

these regions.

When the gradient wind is from higher pressure lakeward the diurnal pulsing of the lake high serves mainly to reinforce this flow. The vertical extent is much less than for the opposed parallel wind condition. The only cloud feature as a direct consequence of this gradient wind may be a line of cumulus inland at the convergence zone. Again, this cloud line will mark the limit of the lake air movement.

THE LAKE ANTICYCLONE AND AIR POLLUTION

One of the immediate consequences of springtime lake effect is the ability of the converging or diverging winds to concentrate or disperse aerosols near the shoreline. With the heavy industrial area of Chicago, Hammond, and Gary along the southern shore of Lake Michigan some surprising results of thunderstorm increases have already been uncovered by Changnon (1968). Excess precipitation appears to be related to increasing convective activity and condensation nuclei.

The following table (Table 5) presents the visibility at 1300E for Waukegan and over the lake during the observation period. These observations are classified by gradient winds as was done in the preceding chapter. The increased incidence of reduced visibilities at Waukegan under southerly winds is highly reflective of the Chicago industrial area's contribution of air pollutants. In all four cases, visibilities were reduced both over the lake and over the land. One day, 25 May 1967, saw the development of a lake high of sufficient strength to restrict the visibilities more within the suppressed vertical motion zone over the land. This circulation permitted anomalously higher visibilities over the lake. The lake high is most effective in dispersing atmospheric pollution under an onshore gradient wind (i.e. Easterly) as this flow carries stack effluent landward where turbulent processes maximize the aerosol diffusion.

A weak offshore gradient wind that allows a lake breeze to develop will produce an "atmospheric stack" at the convergence zone. Pollutants emitted into the convergent flow on either side of the lake breeze front will be lifted at the "front" and then carried lakeward to become diffused over the lake. Some of these aerosols undoubtedly become trapped

TABLE 5. R/V MYSIS and Waukegan visibility under different gradient wind directions.

Gradient Wind	Date	VISIBILITY Waukegan	Mysis
Northerly	24 May 1967	15 km	9 km
Easterly	1 June 1967	28	18
Easterly	2 June 1967	28	18
Easterly	3 June 1967	28	18
Southerly	25 May 1967	7	9
Southerly	4 June 1967	15	2
Southerly	5 June 1967	9	7
Southerly	26 May 1967	9	---
Westerly	8 June 1967	28^+	---
Westerly	25 May 1966	---	28^+

by subsidence contributing to the haze within the inversion over the lake. The typical visibility report from ships during spring is "visibility reduced to 9 kilometers by haze." Winds paralleling the shore will form a band of lowered visibility downwind from the source. Opposing pressure gradients of the general wind and the local lake high and the more extreme vertical mixing, would aid in dispersing aerosols more effectively than the gradient wind which results from higher pressure lakeward.

As an illustration of this problem: If Chicago represents a pollution source, a southerly spring or early summer wind will not have as much dispersion capabilities as will the northerly wind. In another light, if both Milwaukee and Chicago (along the west shore of Lake Michigan) emitted equal quantities of aerosols the reduction in visibility at Milwaukee, when the wind flow was from Chicago, would exceed the reduction noticed at Chicago during reversed wind flow.

THE CHANGING LAKE BENEATH THE ANTICYCLONE

General temperature characteristics and water movements

By early April the ice is nearly gone from Lake Michigan. Water is nearly isothermal at considerably below 3.9°C both vertically and horizontally. During the remainder of April, warming in response to the increasing insolation characterizes the lake. Warmer air temperatures and rains directly increase the heat input to the lake and add it also via river runoff. Until maximum density (3.9°C) is reached, the warming surface water becomes more dense and sinks, undergoing vertical mixing (overturn) with the entire water column. In extremely deep areas of Lake Michigan this process may continue well into the month of June. In 1967 at the "deep hole" (282 meters) off Frankfort, Michigan, overturn continued until early July.

Shallow nearshore waters quickly saturate their water columns with heat to and above maximum density for two reasons: 1) The shallow depth requires less input of total heat energy before the water column warms to maximum density, and 2) the influx from rivers and beach drainage indirectly contribute additional heat nearshore.

The thermal picture of the surface waters of the Great Lakes shows a surface gradient nearly parallel to the basin bathymetry during the months of May and June. Basically this surface pattern consists of warm shallow waters and cold mid-lake waters separated by a developing temperature gradient.

Neglecting, for the present, any other forces that may be at work at the surface, the result of this horizontal surface temperature gradient should produce a circulation within the surface waters. Stokes (1849)

and Thompson (1869) were the first to theorize that thermal gradients could drive a circulation. This baroclinic state was elaborated upon later in the nineteenth century by Bjerknes (1898). The results from the efforts of these men were the establishment of the hydrostatic and geostrophic equations. As the shallow water warms more rapidly along the shoreline (above 3.9°C) the increasingly less dense water becomes slightly higher in its surface elevation as it expands in warming. The increase in elevation represents a "dynamic height." The slope of the surface from warm to colder mid-lake water becomes a direct function of the thermal energy contained in the water column. With the addition of a rotational coordinate system (the Earth) Coriolis force adjusts this pressure gradient force, caused by the dynamic height, to form a geostrophic current. The results of the generalized closed-current system of a lake in spring are shown in Figure 31. Since the center lake water is at maximum density, no thermal bar needs to be represented.

<u>Influence of the stability on the wind stress at the interface</u>

The wind driven circulation is a much debated subject among oceanographers. The air-sea interaction phenomenon of momentum flux is present when the atmosphere is of a neutral or negative stability, however this interaction becomes reduced under positive stabilities. Edmund Halley made the original reference to this coupling in 1686. Since that time Maury (1855), Ekman (1905), Sverdrup (1947), Stommel (1948), Munk (1950), Miles (1957), and many others have added theory and physical interpretation to this complex problem. Energy clearly can be transferred from air to water where resulting motion, both in waves and current, is produced. A quantitative relationship has yet to be resolved. The picture has become even more complex in recent years through the considerations

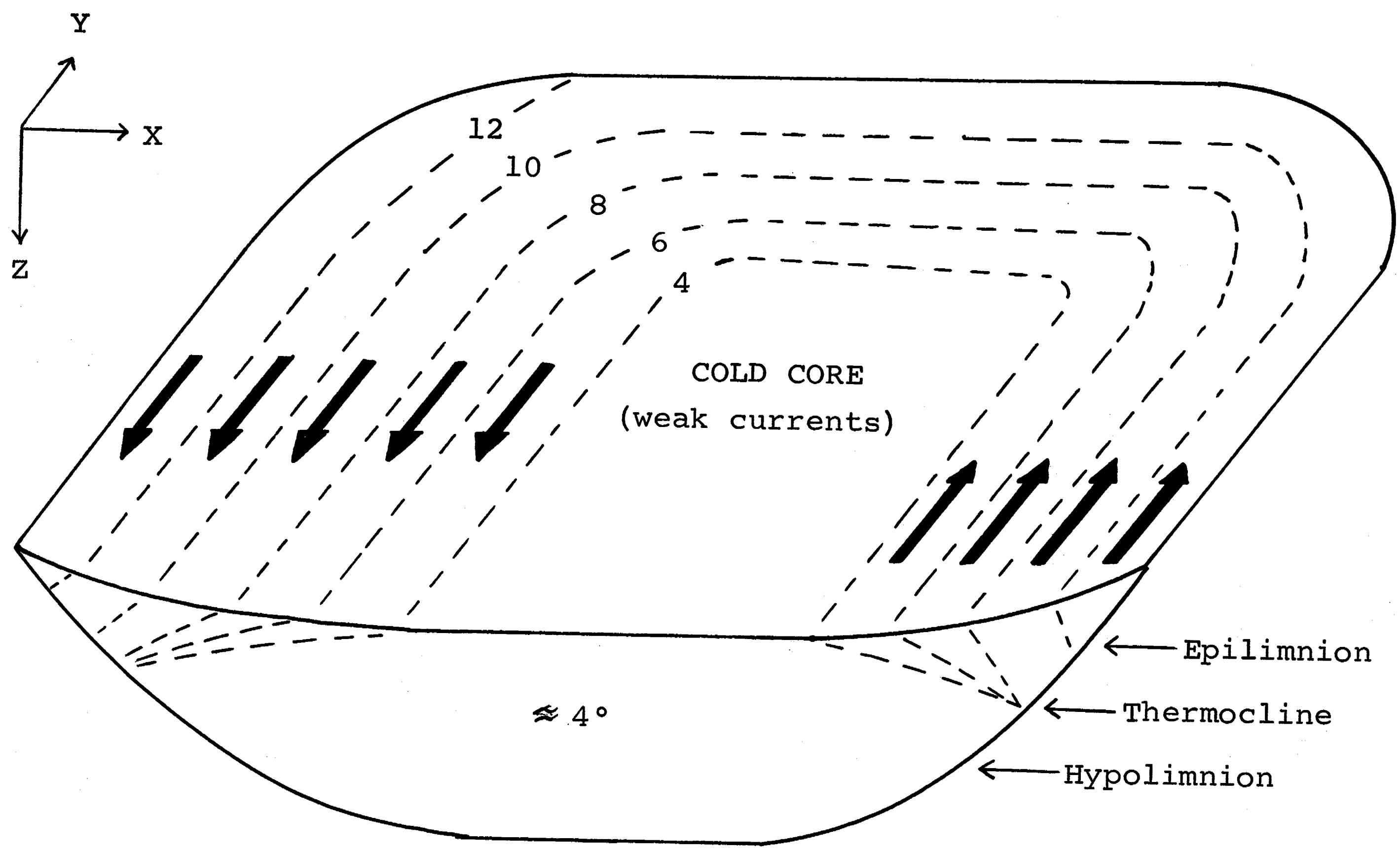

FIG. 31. Three dimensional closed-current system of a warming lake. Vertical height exaggerated. Surface water temperatures in °C.

of Stern (1965) and the geostrophic vortex. The intensity of any circulation vorticity appears to respond to the wind stress at the interface in the production of both vertical and horizontal currents.

When the complications of a stable atmosphere overlying the water are included, the momentum transfer process becomes even more complex. Vertical motions characteristic of the turbulent atmosphere are enhanced by instability and reduced by stability. The relation of these differences to wave generation capabilities have been investigated by Strong and Bellaire (1965), Jacobs (1965), Cole (1967), and Richards et al. (1966) for the Great Lakes. As noted above, substantial air-water temperature differences may prevail over much of the year on the Great Lakes. These differences result from a lag of nearly one month in the annual water temperature cycle behind the air temperature over land upwind from the lakes. Average annual air and water temperatures for Chicago are shown in Figure 32. During spring and fall the temperature difference (T_{air} - T_{water}) becomes most pronounced. As a consequence of the warm water during fall and winter the Great Lakes frequently create a low pressure area (Petterssen and Calabrese 1959) and act to attract low pressure centers moving across the United States (Strong 1968). This increase in storm track frequency has also been demonstrated by Klein (1956-1957).

Much the opposite effect predominates during spring. The strong stabilities, resulting from the lingering cold lake water, produce an increased frequency of anticyclonic weather. Klein has also demonstrated the favored anticyclonic path to be across the Great Lakes region throughout May and June. Since it is with the spring period that this paper is concerned, an attempt is made to physically interpret the various means of energy transfer into a cold lake.

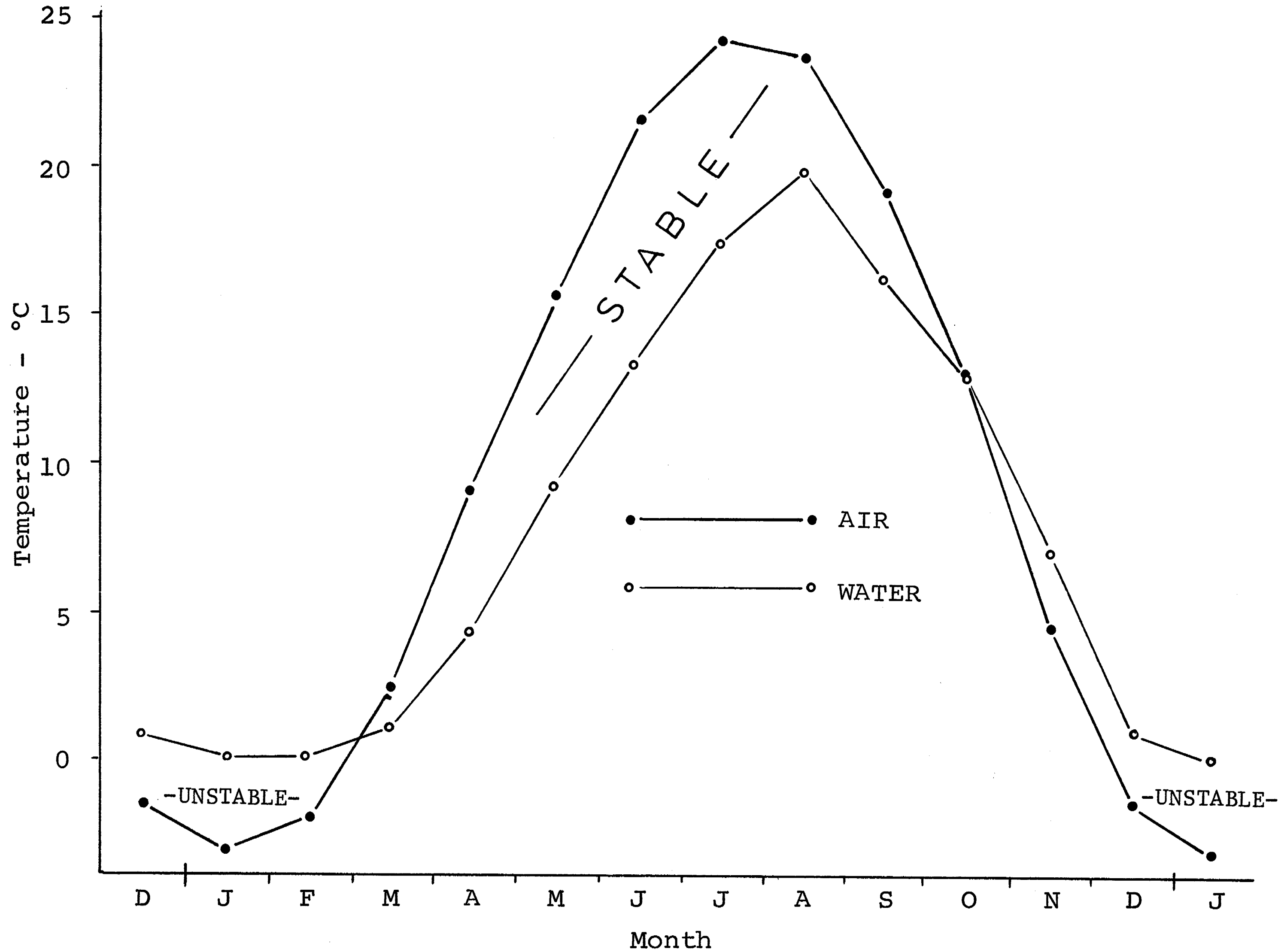

FIG. 32. Chicago average air (Midway) and water (Intake cribs) temperatures by month - 30 year averages (1931-60).

Figure 32 indicates the high frequency of stable days that prevails over Lake Michigan during spring. Calculation of Richardson number profiles have been attempted on various occasions and in varying localities. Data obtained from a meteorological tower 1.6 kilometers from the eastern shore in Lake Michigan (Elder 1964) have been analyzed for inversion conditions. On 15 September 1967 an abnormally warm offshore wind (southeasterly) produced a strong inversion. The average Ri profile for the afternoon (1330E to 2030E) is presented in Figure 33. The wind profile was nearly linear, similar to what Liljequist (1955) observed over the Antarctic snow field under inversion conditions. The inflection of the wind velocity profile just above the water surface is attributed to the air-water boundary layer. Below this inflection the wind velocity decreased more rapidly to the surface. The air temperature profile is quite similar to inversion measurements reported in Hill (1962). As with the wind profile, the temperature decreased rapidly approaching the the air-water interface (approximately 18°C).

The initial rapid increase in Ri with height and its peak at about 4 meters is similar to the profiles presented for Hansen's (1966) "undulant" condition (see also Hansen and Lang 1967). Hansen showed a maximum Ri just above 10 meters. The differences in flow characteristics over the scrub vegetation at his White Sands, New Mexico, site from that over relatively flat Lake Michigan may account for the lower height of the observed peak in Figure 33.

An average Ri profile has been computed from Elder's wind and temperature measurements (1027 cm, 699 cm, 423 cm, 227 cm, and 100 cm above the surface) which were made once every two minutes over a period of seven hours from 1330E to 2030E (210 periods of data) for 15 September

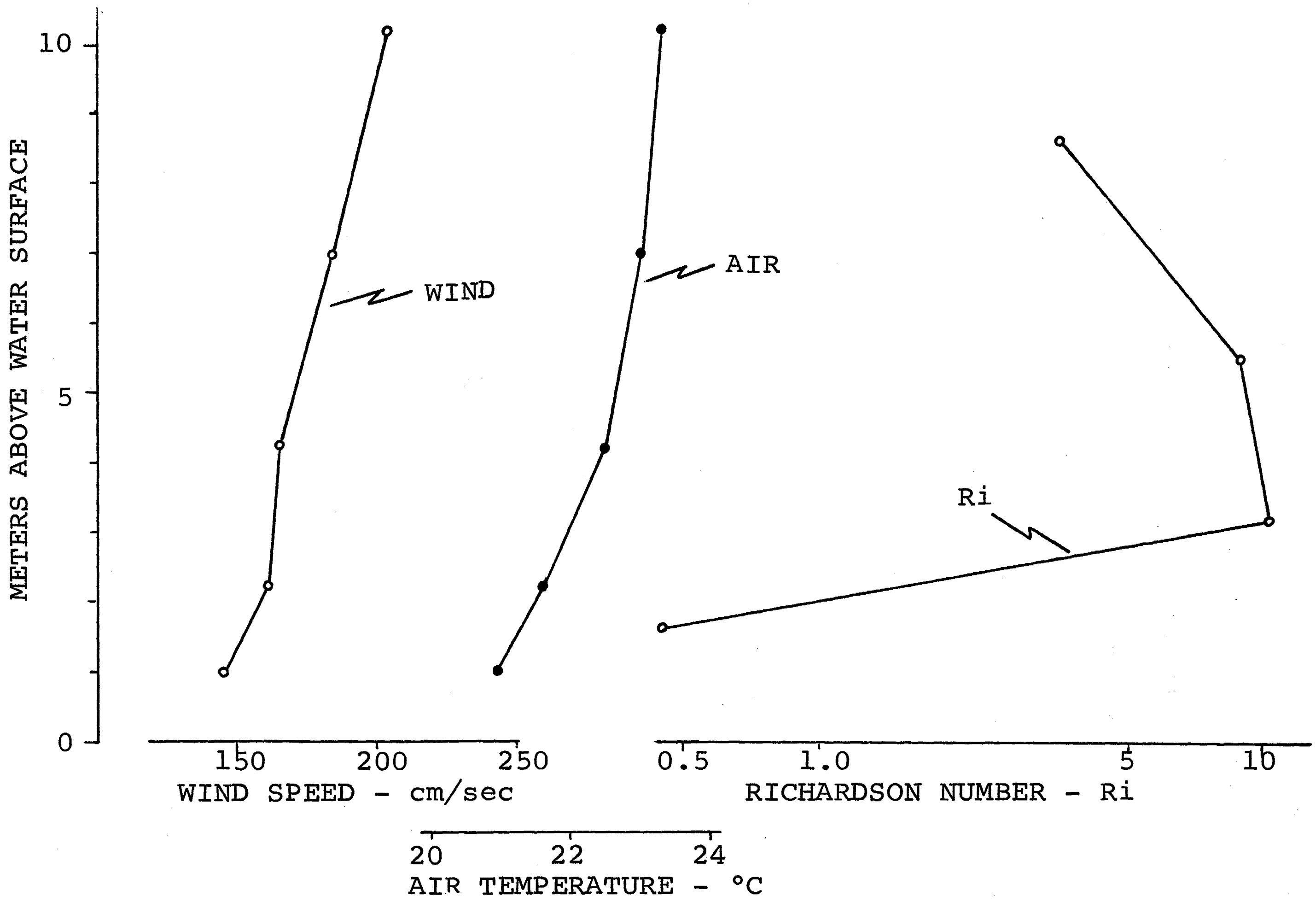

FIG. 33. Profiles of wind, air temperature, and Richardson Number for 15 September 1967 at tower 1.6 km off Lake Michigan shore.

1967. This is the same data that was averaged for Figure 33. These data, averaged hourly, are shown in Figure 34. The highest Richardson value observed during this period was +48, and is indicated. This peak occurred just prior to sunset between 423 and 699 cm over the water surface.

A critical Ri for heights close to the earth's surface has been sought for by many authors. This critical region, it is argued, occurs where eddy viscosity becomes relatively less overwhelming in comparison to molecular viscosity. With increasing values of positive Richardson number the flow becomes less turbulent. Richardson (1920) proposed that this critical region was in the vicinity of Ri = +1. Schlichting (Liljequist 1955) has indicated +1/24 may be more probable. Recent scintillation data over snow (Portman et al. 1962) indicate a break in the curve relating percent modulation of a constant light source per layer temperature difference [%mod./$(T_2 - T_1)$] to Ri at 1.5 m. They found a change in relationship to occur at +0.35. McVehil (1964) investigated the relationship of Ri and Deacon number (β). He used data from both over grass at O'Neill, Nebraska, and over snow (South Pole) measurements. The relationship in positive Ri regimes was found to be linear and inversely proportional until about Ri = 0.1 above which wide scatter, much the same as with the scintillation data, becomes evident. If the linear relationship be empirically extended to $\beta = 0$ (when it could be hypothesized that no wind shear occurs with height) the Ri value obtained for such a condition is +1/7. McVehil further indicated that the ratio of heat flux to momentum flux coefficients, K_H/K_M, which is close to unity under neutral conditions, decreases with increasing stability. The implication from observations of inversions is that K_M is

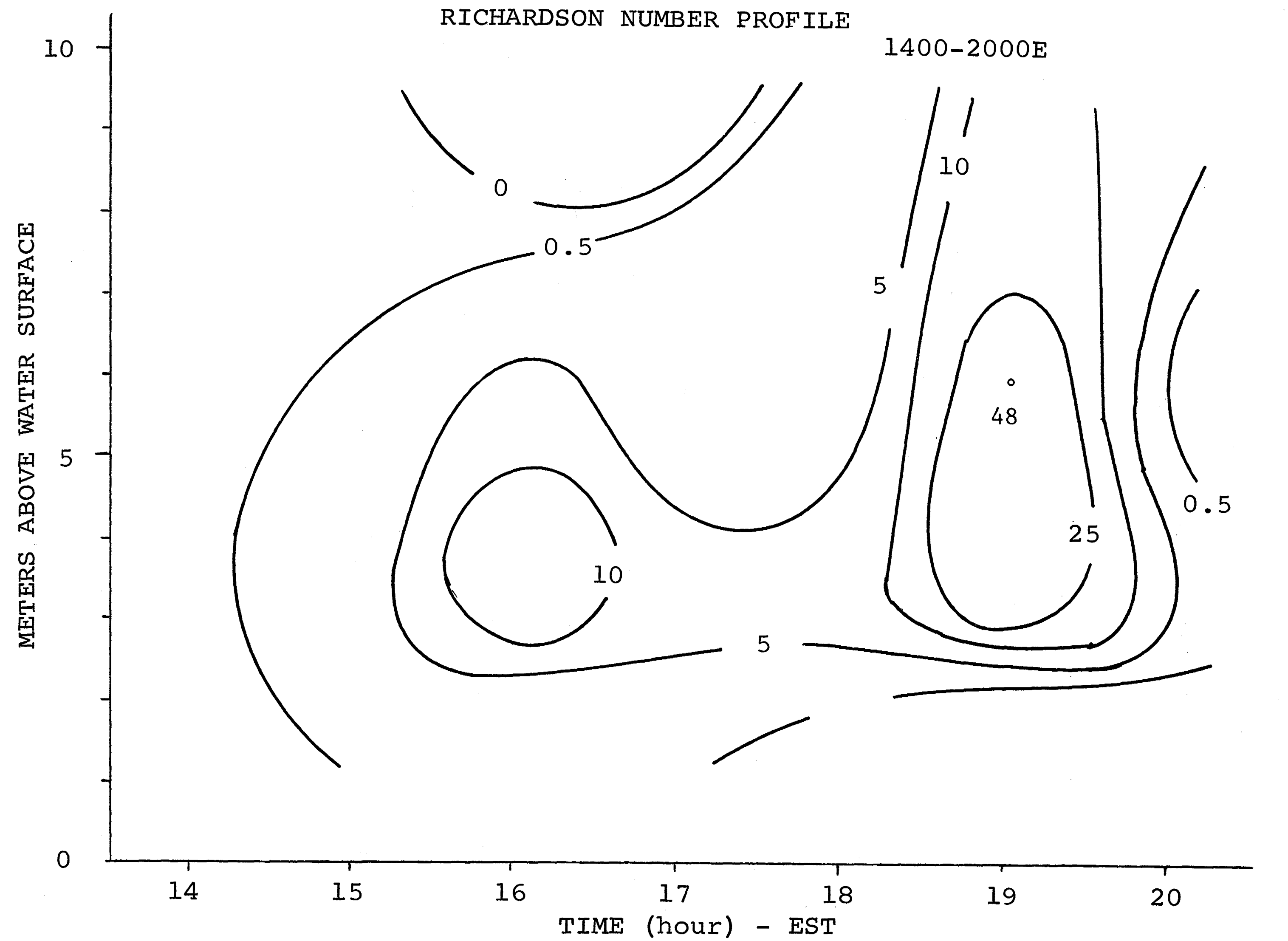

FIG. 34. Richardson Number profile at the Lake Michigan research tower on 15 September 1967.

effectively reduced but that K_H is reduced to an even greater degree. If K_M approaches zero with increasing stability, K_H must approach zero more rapidly.

The formal definition of any vertical flux, F_s, may be written:

$$F_s = \rho K_s \frac{\partial s}{\partial z}$$

where K_s = the eddy transfer coefficient of an entity "s." Hence, the vertical flux of any conservative entity, whether it be water vapor, momentum, or sensible heat energy, is a function of the vertical gradient of that entity. For each characteristic flux an eddy coeffieient (K_s) exists which is in turn a function of the turbulent condition of the atmosphere. The eddy coefficient is also inversely dependent upon the height above the surface. Obukhov (1946) has demonstrated that for near neutral conditions: $K_M \simeq K_H$. For the inversion, this regime of apparent decreasing transfer downward as the surface is approached seriously affected Estoque's development of his numerical sea breeze. His lower 50 meter layer of constant flux seriously hindered proper development of the sea effect (lake effect) winds in the lowermost region of his model.

McVehil (1964) has shown that for Ri > +0.08 values of K_H/K_M display an apparent decrease from unity. This decrease has been explained by Townsend (1958) and Stewart (1959) by gravity waves operating on the inversion. The inversion, they claim, serves as an inhibitor to heat fluxes while momentum transfer is partially maintained through pressure oscillations set up by the gravity waves. Hill (1962) has indicated an average value of $K_H/K_M \simeq +0.7$ for the Ri interval $0 \leq Ri \leq +0.08$. No values have been related to more positive stability conditions where the

flow approaches an increasingly less turbulent state. If Townsend's and Stewart's interpretation is valid, the indication would be that this flux ratio is less than +0.7.

The spring-associated anticyclone of the Great Lakes is basically an isolated feature which is gradually acted upon by two major forces. The sun continually radiates energy into the water and eventually develops the thermocline, separating the surface and deep waters into two density media. This separation allows the less dense surface waters to heat more rapidly. The second influence on the warming process is provided by the regional pressure gradient through its interaction with the local lake high. While little heat is transported through the inversion, momentum flux although reduced can, upon occasion, mix the heat in the surface waters to deeper levels within the lake. This mixing serves to intensify the developing thermocline.

As the edges of the lake warm, more interaction of air and water is permitted there and a hybrid model develops. The two outside influences gradually diminish the extent and intensity of the lake anticyclone, at the surface and around the edges. By August the average land air temperature begins to decrease and direct air-water coupling develops.

THE RESULTS OF AIR-WATER INTERACTION UNDER HIGH STABILITY

Surface water temperature during the observation program

The surface water temperature for three periods coinciding with the Waukegan study is presented in Figures 35, 36, and 37. Data over a period of three days have been included in the first map to develop the best possible pattern to which the others may be compared. Figure 35 is a composite of observed water temperatures made by the C/F (car ferry) MADISON (plying from Milwaukee to Muskegon), the R/V INLAND SEAS, and the R/V MYSIS. Temperatures about at maximum density were present over the deepest part of the southern basin of Lake Michigan. The thermal gradient showed some latitudinal effect in the southernmost waters. By the last day of May 1967 (Figure 36) the region of maximum density surface water had shrunk, but the major gradient was still confined to the nearshore areas.

Figure 37, eight days later, shows 6°C surface water remaining in the center of the southern basin. The thermocline is now present across the entire lake as the surface water is all above maximum density. Note the anomalous warm streaks along the Wisconsin shore. These streaks are believed to have resulted from air-water interactions and will be discussed in more detail in a following section.

The lake anticyclone and the lake circulation

If no prevailing wind gradient existed during the spring over the Great Lakes, cloudless days would presumably produce lake breezes similar to those described and presented by Moroz (1967). Such a local wind would be, by its nature, diurnal and altered during its daily existence by Coriolis force. In Figure 38 the basic, Coriolis-compensated, anticyclonic

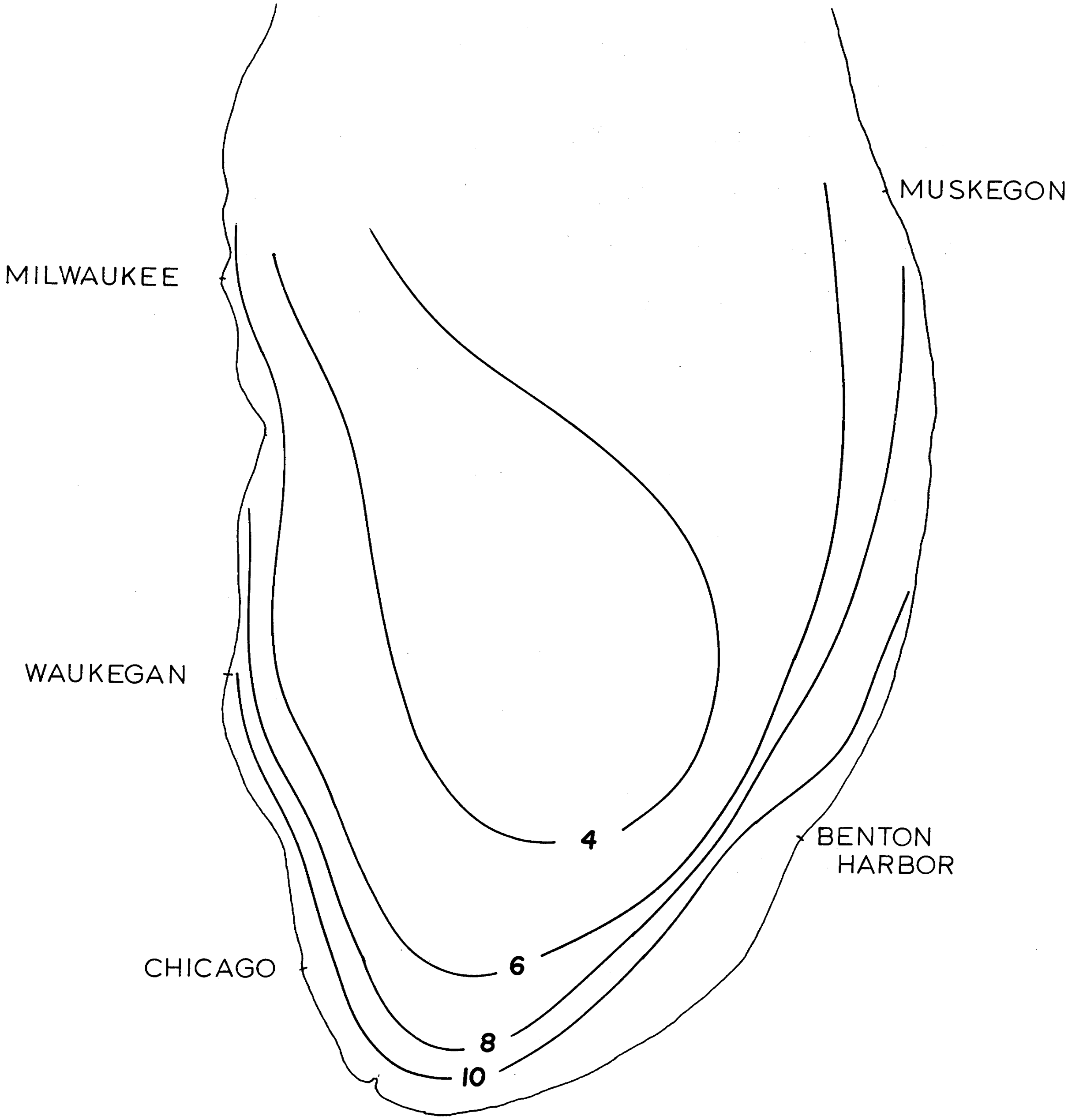

FIG. 35. Lake Michigan surface water temperatures (°C) 22-24 May 1967.

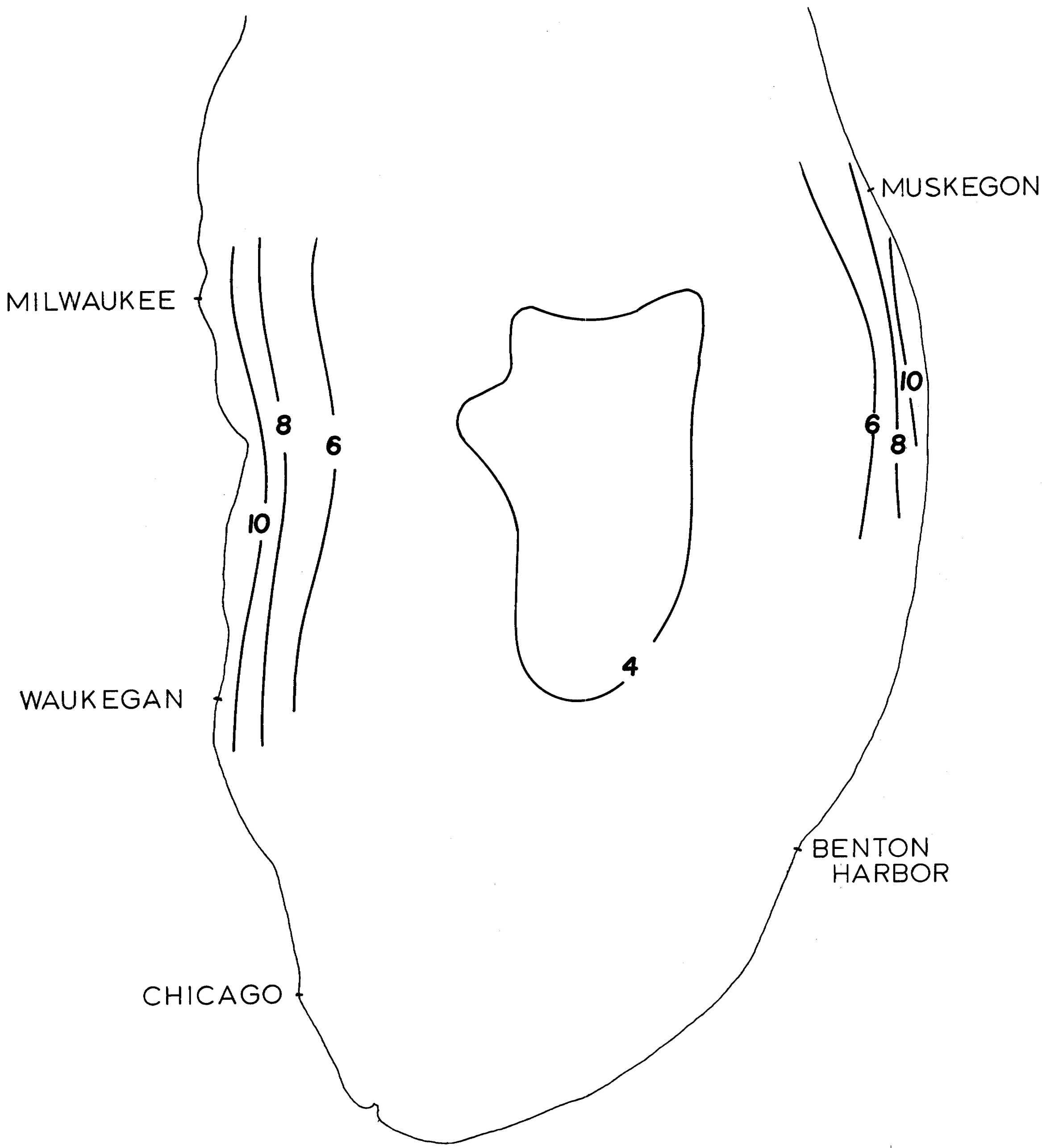

FIG. 36. Lake Michigan surface water temperatures (°C) 31 May 1967.

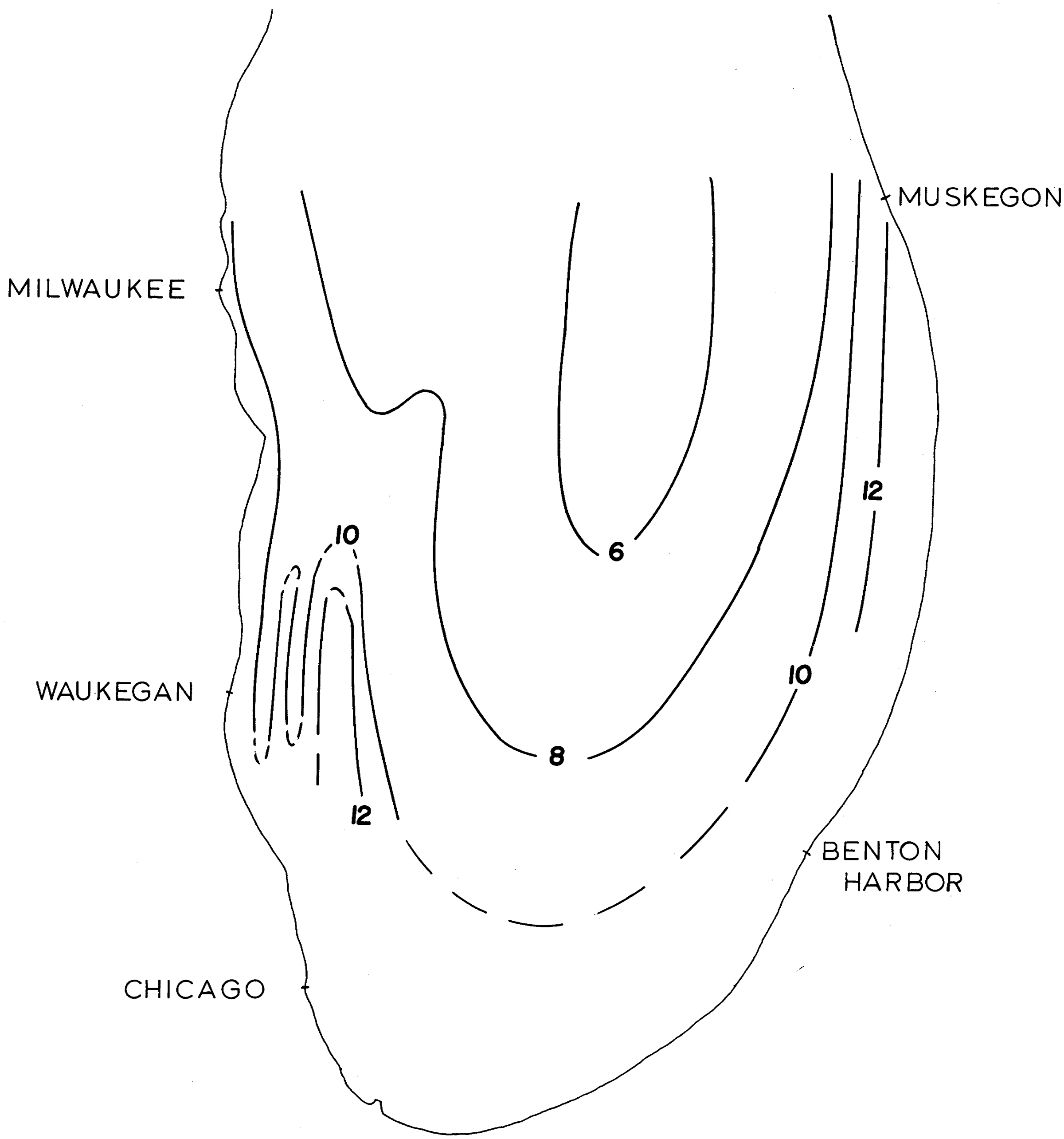

FIG. 37. Lake Michigan surface water temperatures (°C) 8 June 1967.

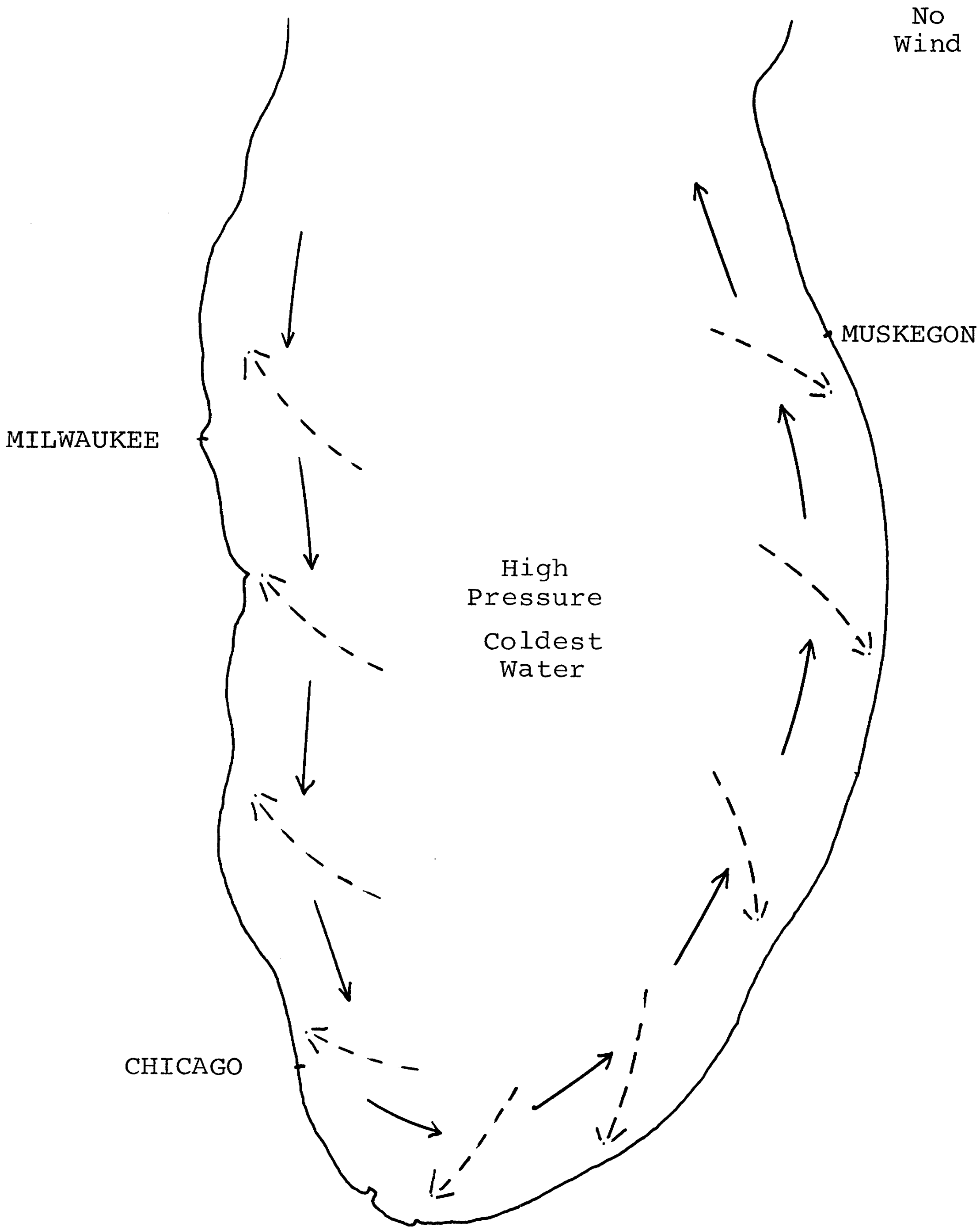

FIG. 38. Lake breeze (zero gradient wind) surface wind stress (dashed lines) and inhibited thermal current (single line arrow).

surface winds (dashed arrows) of the pure lake high have been superimposed on the basic cyclonic thermal current circulation (solid arrows) for spring. These surface winds indicate the direction of an applied wind stress to the water surface. While the atmospheric and water circulations oppose one another, the low level subsidence and divergence in the atmosphere of the lake high tends to support a slight upward vertical circulation within the mid-lake waters where no thermal circulation is present. This slow upwelling carries warming surface water away from the lake center. The thermal circulation of the water is at no point reinforced by the wind stress so that any cyclonic mass transport during a pure lake high condition would be inhibited. The lake high becomes a third important factor in maintaining the distribution of lake heat as the upwelling it produces has a tendency to restrict the warm buoyant surface waters to the nearshore areas. The pulsing effect of the diurnally vacillating lake high, as it mixes and restricts the surface water landward, can be seen in the water temperature summary for the Waukegan coastal region in Figure 39. This effect seem most pronounced on the first few days in June when the lake high was most intense.

The perturbation of the lake anticyclone by the gradient wind

While the stable lake anticyclone considerably limits momentum transfer to the water during spring, a coupling is provided through the top of the inversion. The perturbation of the lake anticyclone by the gradient wind may be analogous to the wind stress coupling with the geostrophic vortex (Stern 1965) represented by the thermal water circulation. This analogy is demonstrated by the apparent vorticity increase (in the negative sense) that the lake anticyclone displays. The inversion merely acts as a secondary interface. The three fluids become coupled

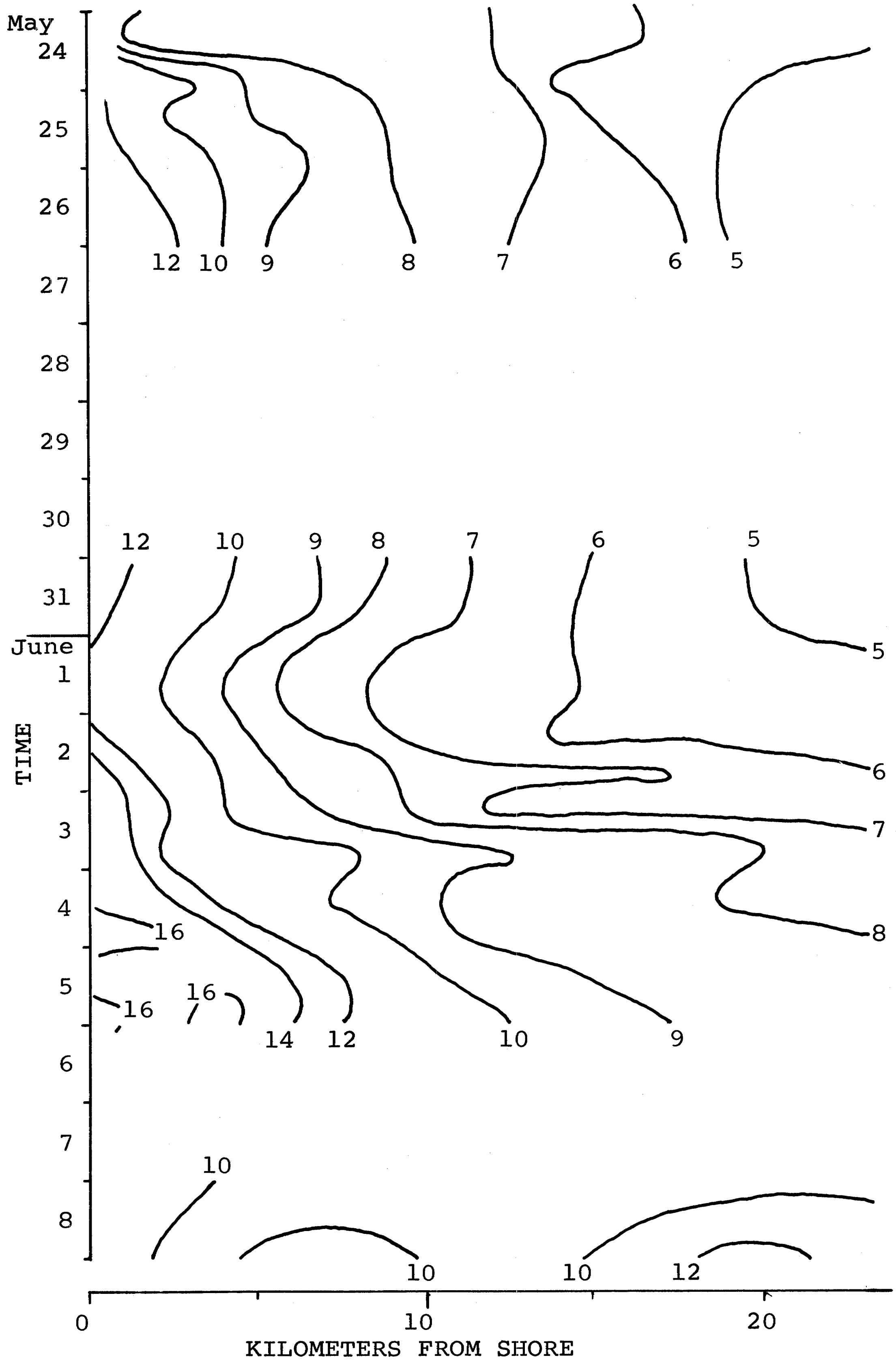

FIG. 39. Observational period inventory of surface water temperature (°C) off the coast of Waukegan, Illinois. 1967.

through shearing forces of the warm air-cool air interface and the cool air-cold water interface. Therefore, the efficiency of the geostrophic wind in both producing waves and conveying sensible heat is minimized as turbulent energy transfer is inhibited by the additional interface.

Perturbation of the water circulation from a basin-oriented lake-edge cyclonic thermal structure becomes a function of prevailing gradient wind. Basic low level wind stress is suggested in the modified Estoque models - see Figures 1 and 2 and our results. A series of figures, 40-43, presents these surface wind stress considerations applied to Lake Michigan for northerly, easterly, southerly, and westerly gradient winds. Thermal circulation is shown to be inhibited where wind stress is opposing the current (single current arrow) and enhanced where wind stress acts in the same direction (double current arrow). Due to the shape of Lake Michigan the southernmost region is highly conducive to eddy formation. To preserve continuity, 1) surface waters must sink when strong surface currents converge into an area of inhibited currents or 2) water upwells when surface currents diverge. The south end of Lake Michigan represents a compromise area for the general water transport along the west shore with the differing character of the east shore transport. The cyclonic thermal circulation incorporates this alteration in transport intensity by an adjusting vertical current. Such a feature is made stable by increasing the fluid vorticity in the form of one or several vortices or eddies. The upwelling eddy will have a cyclonic circulation while the sinking eddy will rotate anticyclonically. Nearly every attempt at current mapping of Lake Michigan shows some eddy feature between Chicago and Benton Harbor in the southern basin - Ayers et al. (1958), Verber (1964), Bellaire (1964), Harleman,

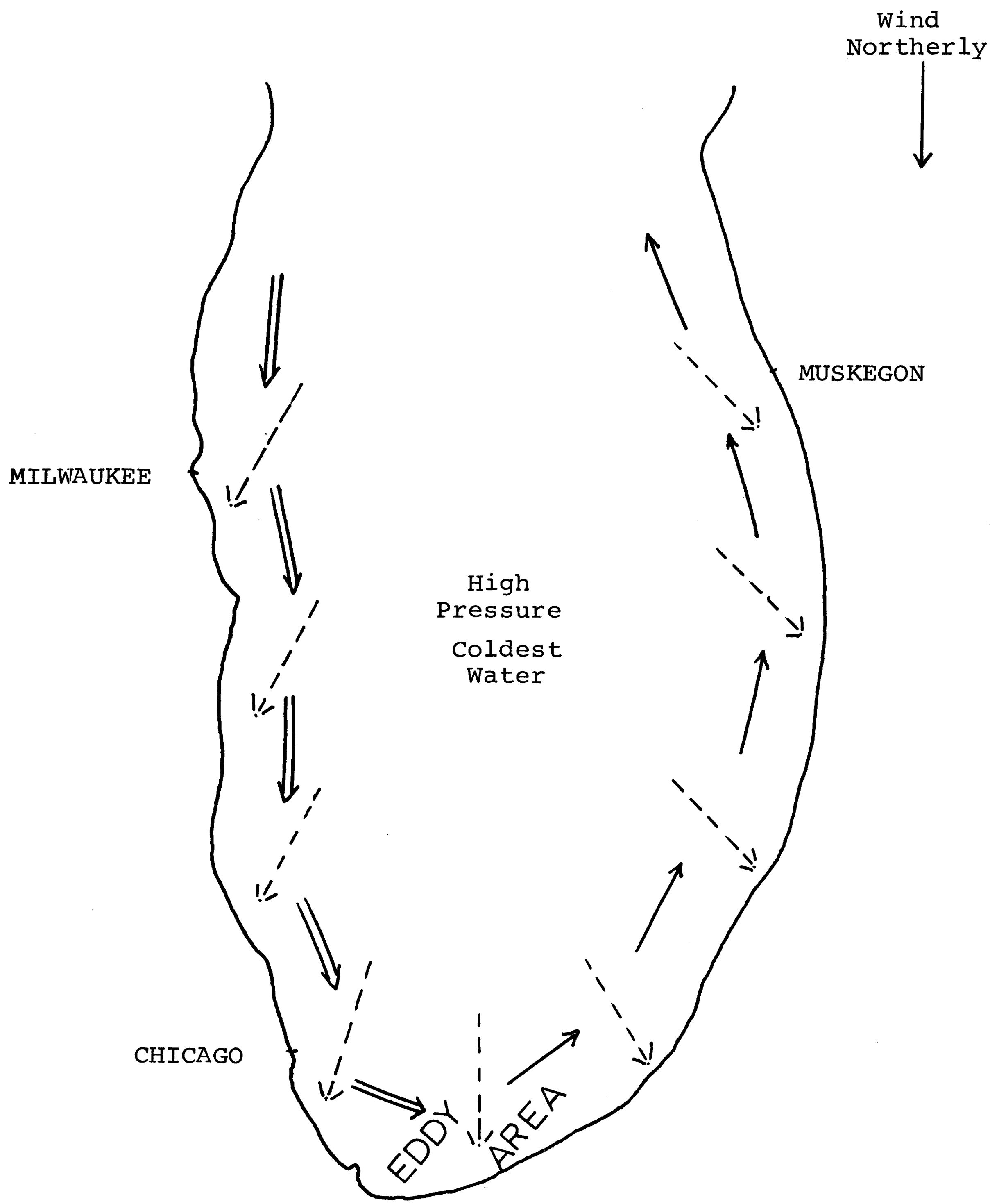

FIG. 40. Northerly wind lake effect. Surface wind stress (dashed arrows). Thermal circulation: single arrow - inhibited current, double arrow - enhanced current.

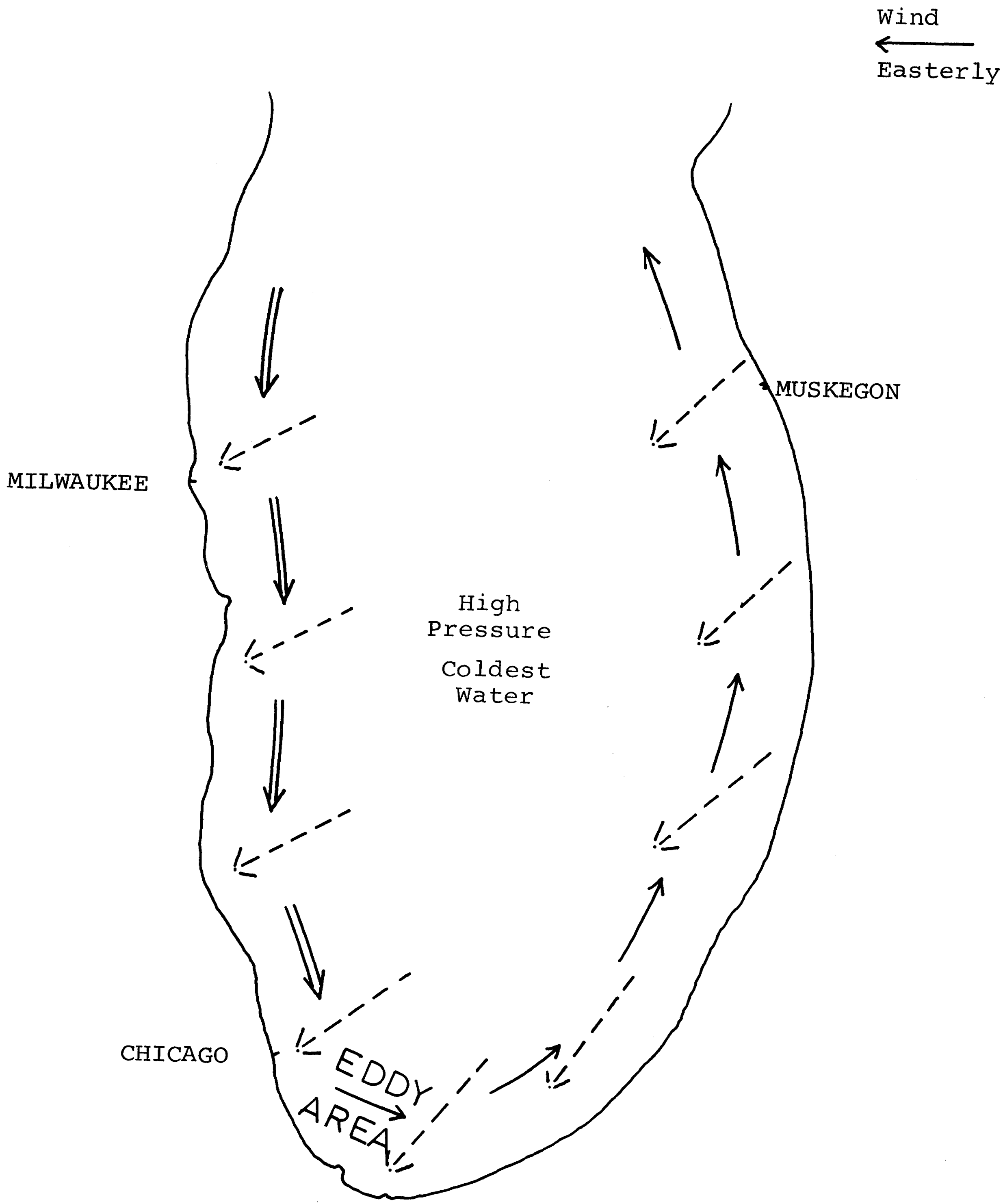

FIG. 41. Easterly wind lake effect. See Fig. 40 for details.

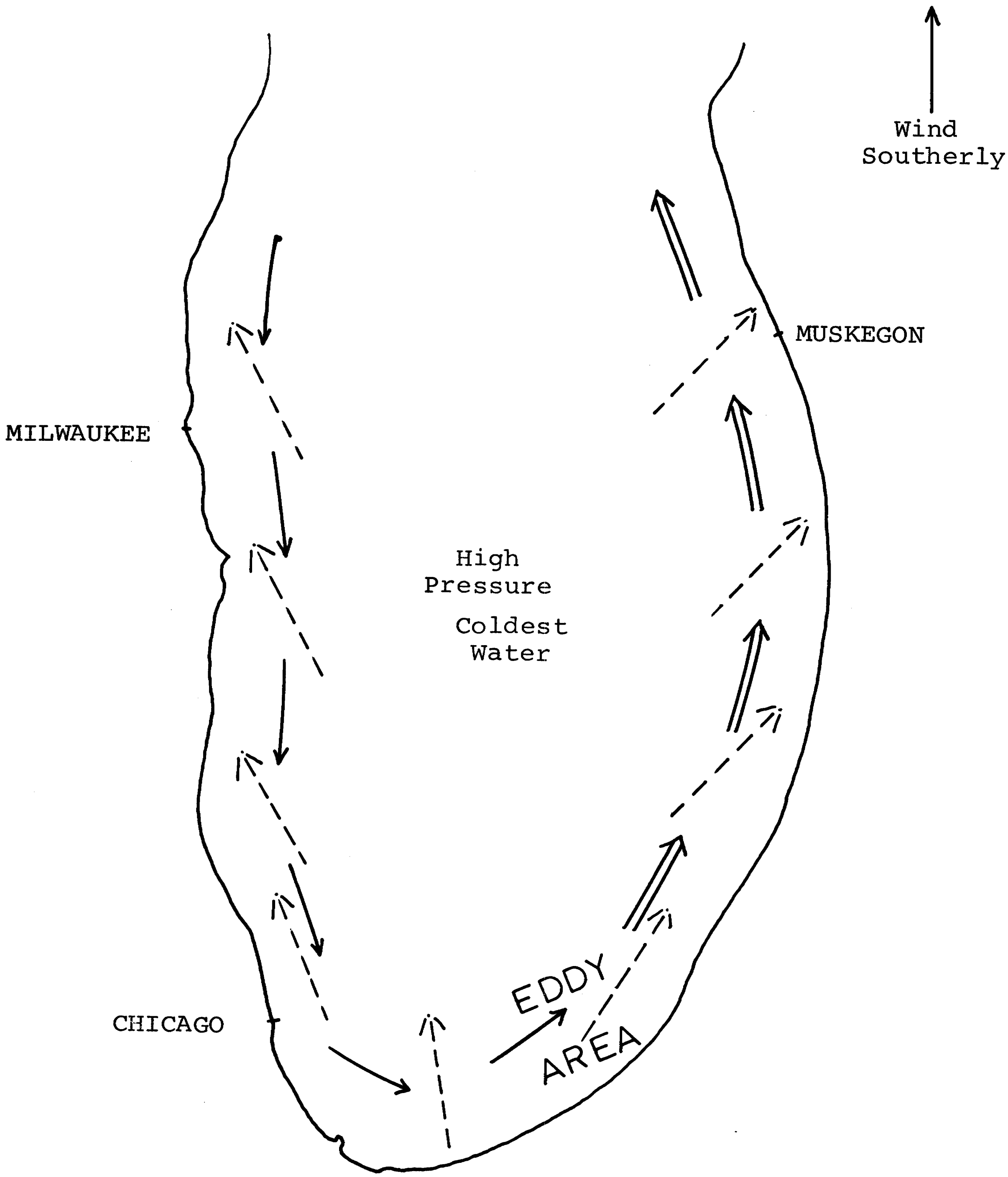

FIG. 42. Southerly wind lake effect. See Fig. 40 for details.

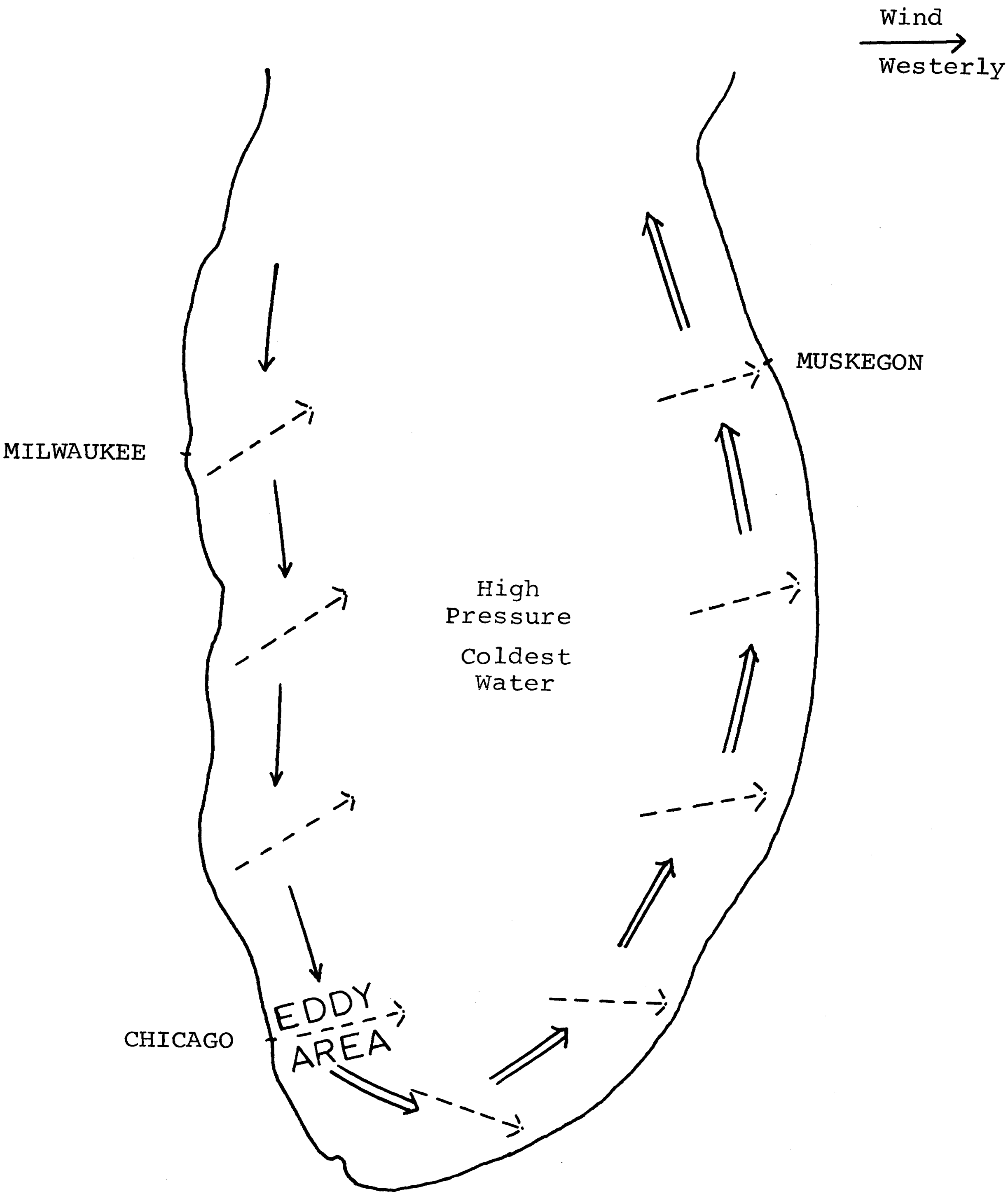

FIG. 43. Westerly wind lake effect. See Fig. 40 for details.

Bunker, and Hall (1964), and more recently, The Federal Water Pollution Control Administration (1967).

Prevailing southerly through westerly wind gradients during spring and summer provide a high frequency of the conditions indicated in Figures 42 and 43. The region of upwelling may explain the anomalous reports during June of high water transparencies between Chicago and Benton Harbor - visibility from the surface to as deep as 15 meters. These conditions have been observed by the ship captains and scientists (data logs) aboard research vessels during the past several years as transparencies over the southern basin, as a rule, have become reduced through eutrophication. To this date no apparent chemical or biological indication has been correlated with this phenomenon.

<u>Current restrictions imposed by the parallel-to-shore gradient wind</u>

While widespread subsidence and low level divergence characterize the central region of the lake anticyclone, a balancing lake effected interaction with the gradient wind typifies the nearshore region under a parallel-to-shore gradient wind. The most extreme interaction dominates the shore where the gradient wind opposes the lake high wind inducement (left-hand shore in Figure 2). The unequal vertical interactions between the shoreline air flows apparently cause a slight migration of the divergent center of the lake high toward the shoreline of more vigorous mixing.

The shore whose gradient wind is that resulting from lower pressure lakeward will receive a wind stress which is co-directional with the thermal current. A steady current can be expected along this shore. Conversely, the opposite shore (where the gradient wind is from higher pressure lakeward) will experience a stress in opposition to the thermal

circulation of the basin. The current should be weaker at this shore.

When these considerations are applied to Lake Michigan, the northerly wind (Figure 40) produces a strong steady southward current along the Wisconsin shore while the flow at the Michigan shore is diffuse and sluggish. The southerly gradient wind (Figure 42) reverses these current characteristics. The southerly gradient wind is more typical of the spring and summer months. As expected, current observations along the eastern shore have shown a persistence of northward water transport during this period. A current meter in operation from May until October 1967 near the Michigan shore just south of Benton Harbor displayed no southward current until August. It appears, in light of present evidence, that during spring and most of summer the mean water transport along the east shore concurs with the cyclonic thermal circulation concept.[6]

The anomalous current induced by the normal-to-shore gradient wind

When a warm wind blows across a cold lake a drastically altered effect appears. As seen in Figure 1 and the observations on 8 June, strong subsidence occurs over water at the upwind shore. The center of maximum low level horizontal divergence consequently migrates toward this shore. In the Estoque model the divergence center appears between 2 and 8 kilometers lakeward from the upwind shore. The major consequence of this wind flow is extreme backing of the low level wind vector, which is most pronounced at the upwind shore, but continues beneath the inversion across the lake. If the gradient wind is sufficiently weak (e.g. 25 May 1966 - Figure A20) continued backing will produce the onshore lake breeze at this upwind shore. The wind stress available to affect

[6]Anticyclonic currents along the edge of the Lake Michigan basin have been witnessed during spring; however, the mean mass transport is cyclonic.

the surface water current provides general opposition to the surface thermal circulation at the upwind shore. Coincidental forces at the downwind shore must produce steady and stronger currents northward (see Figure 43).

The Lake Michigan consequences under an easterly wind will provide a southward current along the western shore but favor a more diffuse current along the eastern shore (see Figure 41).

As a special interface interaction of the normal-to-shore gradient wind a local area of upwelling has been observed. A wave-like thermograph trace has been frequently noticed from the C/F MADISON's intake water temperature record as she left port at this time of year (see Figure 4, Noble 1967). Her operation is strictly between Milwaukee and Muskegon for the Grand Trunk Railroad. The wavy trace was most prominant along the west shore but occasionally was observed at the east shore. A nearshore minimum temperature, a feature of upwelling, occurred generally within 3 kilometers of shore. This minimum was followed by a maximum usually within another 3 kilometers. This phenomenon has also been detected on the records of the Great Lakes Research Division's research vessels. The water temperatures obtained from bucket samples rather sporadically during the Bellaire (1965) wiresonde study are shown in Figure 44. The day was typical of a very warm offshore gradient wind. The air-water interaction under an offshore wind became more apparent with further investigations in 1967.

In Figure 37 one can see warm water streaks that have become oriented parallel to the western shore. They were present only at the upwind lake shore. The actual comparisons of ship wind speed, ship air temperature, and ship measured water temperature taken every 0.25 km can be made from

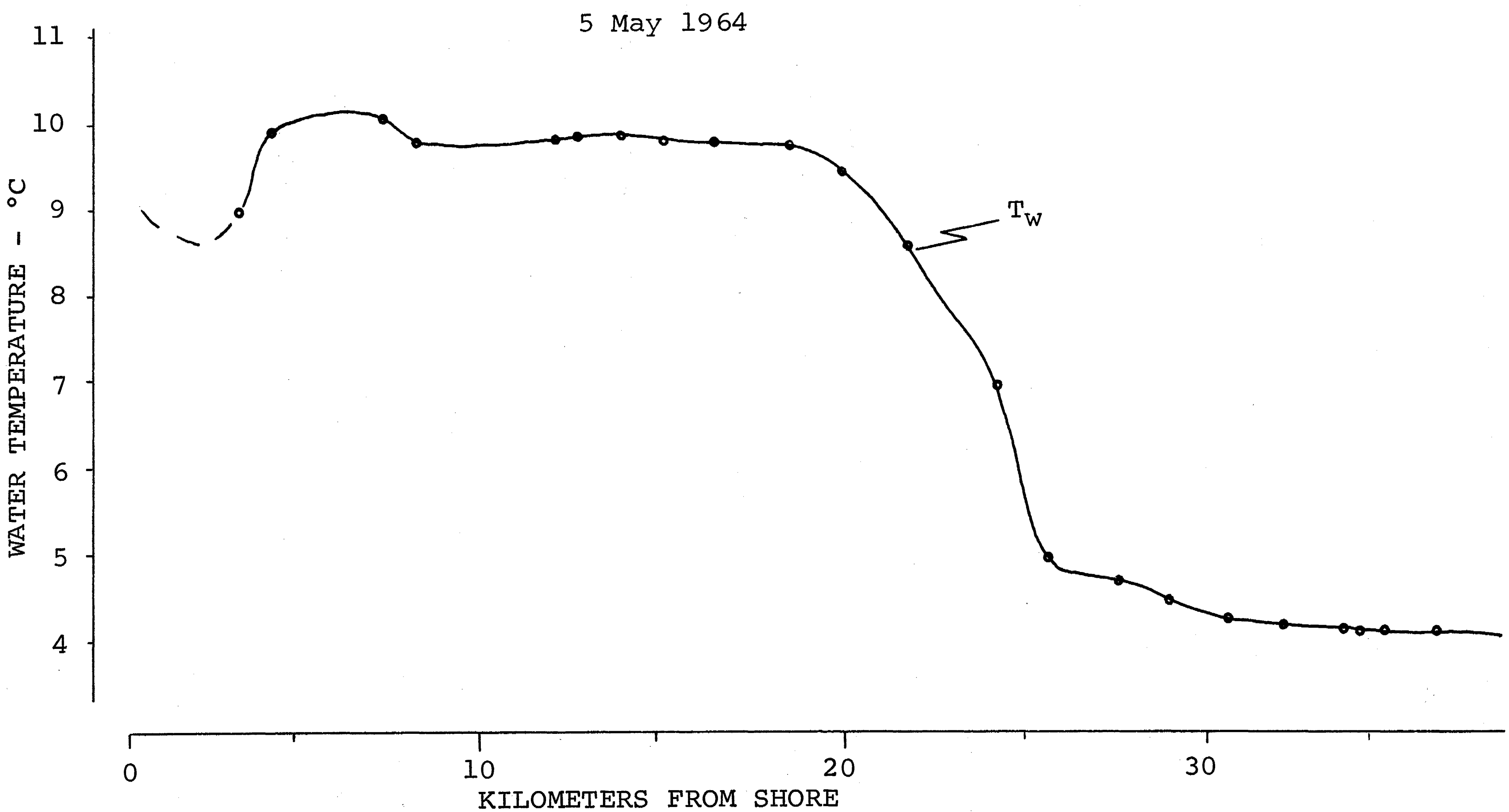

FIG. 44. Water temperature (T_W) profile (°C) perpendicular from shore under strong offshore winds. 5 May 1964.

Figures 45 and 46. Figure 45 demonstrates the warm subsidence and associated cold water upwelling between 0 and 10 kilometers from shore followed by the warmer water streak 15 kilometers from shore and a secondary warm streak 20 kilometers lakeward. These warmer regions are separated by a secondary warm air subsidence region and associated water temperature minimum at about 17 kilometers. Further lakeward, this air-water coupling deteriorates as the inversion becomes more strongly developed over the colder lake waters.

Much the reverse correlations in the atmosphere alone are displayed in Figure 46 between the air temperature and wind velocity at mast level (11 meters). Higher wind speed correlates with higher air temperatures and demonstrates the relatedness frequently noted of K_M and K_H. The atmosphere maintains this related energy transfer further lakeward than does the air-water interaction. The inversion appears to restrict the major air-water interactions within only 20 kilometers of shore under these strong gradient wind conditions as is seen in Figure 45. It is this developing inhibition that protects the lake from turbulent energy transfer toward the central waters of the lake.

Warm water streaks will typify the upwind shore where the wind stress at the surface is in opposition to the thermal gradient of the spring. This condition has been inferred in the Federal Water Pollution Control Administration's report on lake circulation (1967). The streaks appear to be most likely an organized balancing circulation or series of eddy circulations. The wind at this time of year appears not to drive the current directly, but apparently indirectly by transferring energy into the thermally induced circulation or eddies that become generated and maintained by air-water interaction.

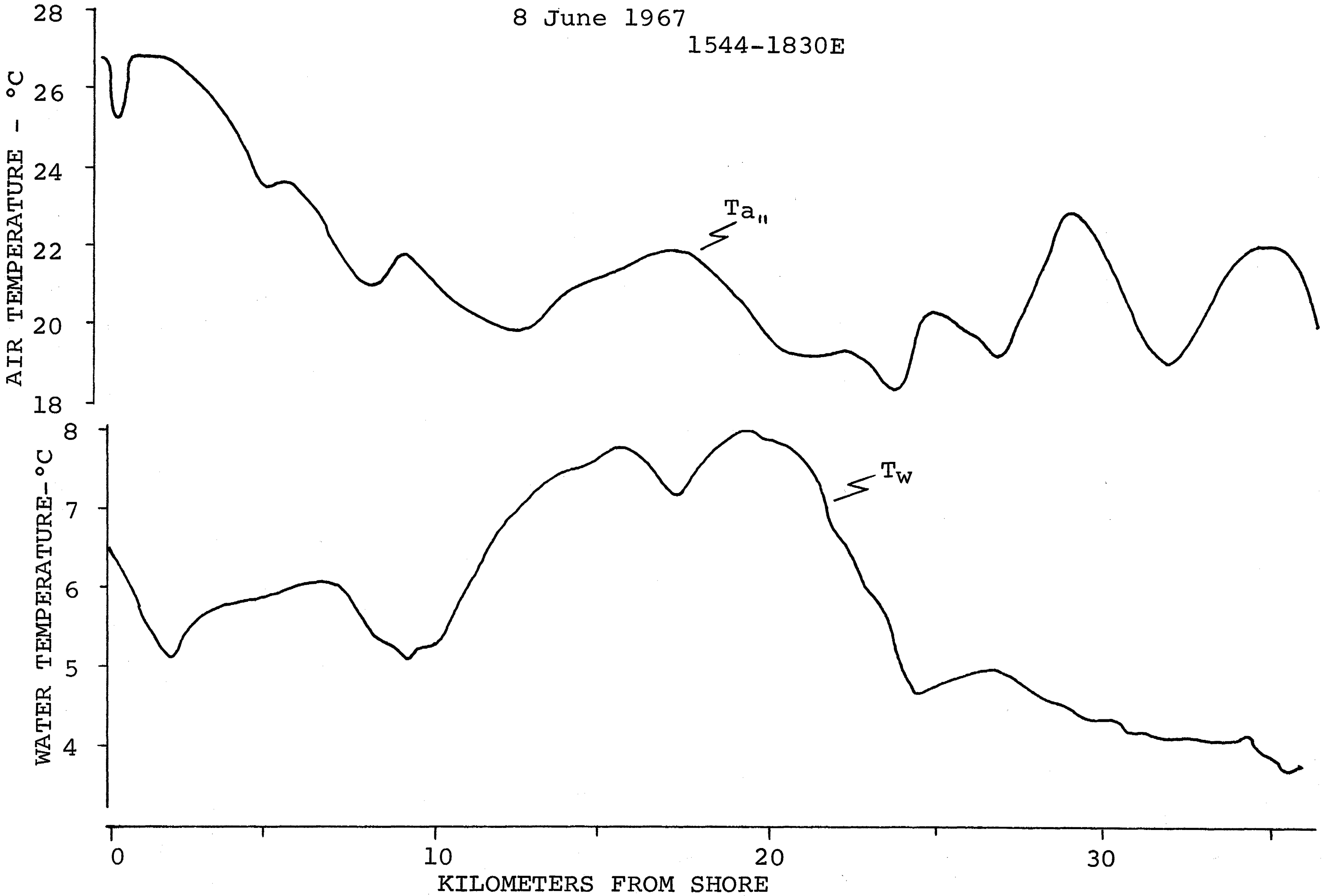

FIG. 45. Surface water (T_W) temperature and 11 m air temperature (T_a) lakeward from Waukegan under strong offshore winds. 8 June 1967.

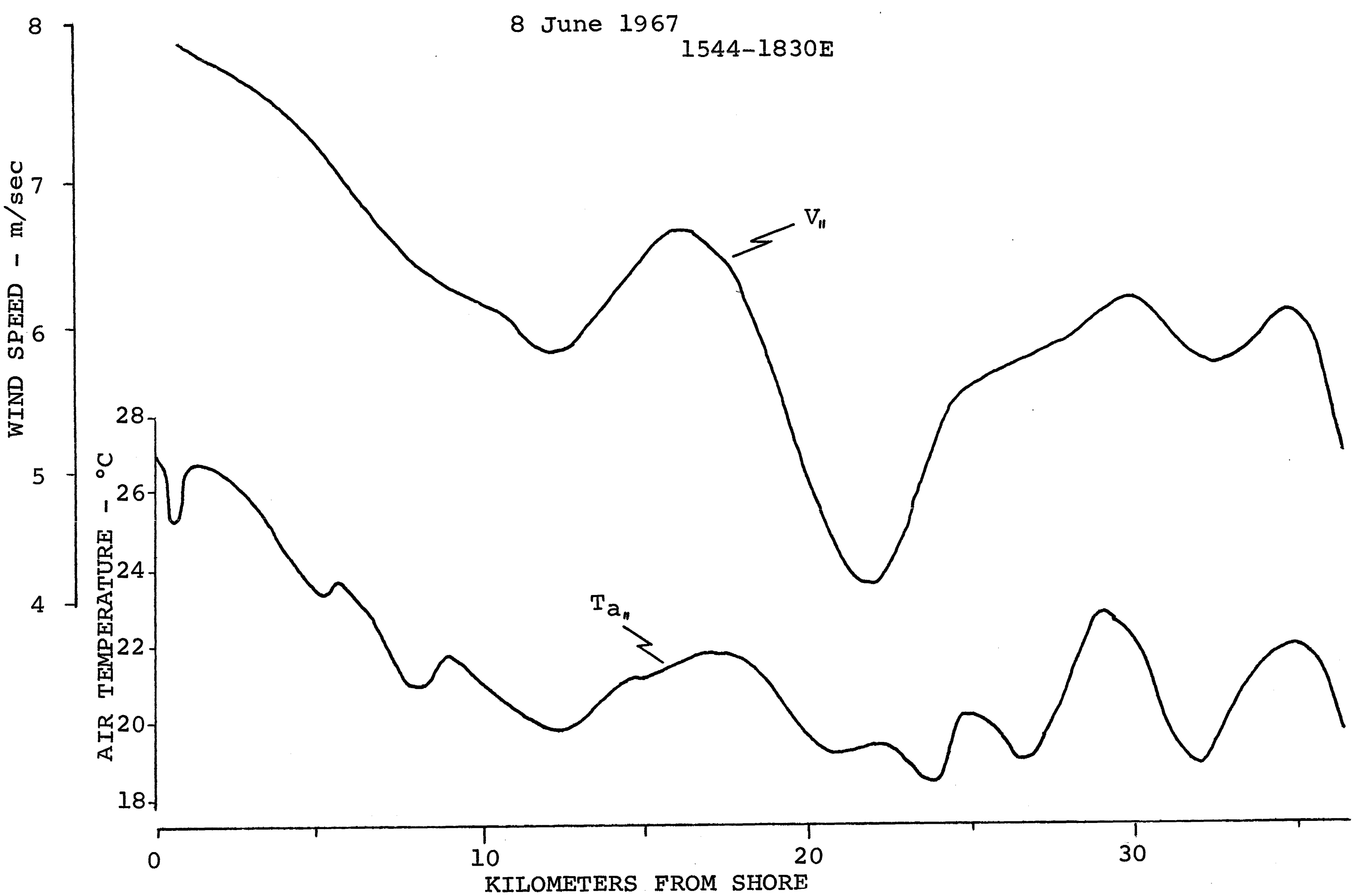

FIG. 46. 11 m air temperature (T_a) and wind velocity lakeward from Waukegan under strong offshore winds. 8 June 1967.

A PHYSICAL MODEL OF A LARGE NORTHERN HEMISPHERE TEMPERATE LAKE DURING SPRING AND ITS EFFECTS ON THE STABLE ATMOSPHERE

The lake effect phenomena resulting from positive air-water temperature differences $[(T_a - T_w) >> 0]$ are a tri-media product. A blanket of air, cooled by conduction to the colder water below, is sandwiched between the general atmospheric circulation of warmer air above and cold lake water below.

Air slowly subsiding through the cold blanket, balances the outflow around the periphery (lakeshore) of this system. This circulation produces a cool high pressure cell with a meso-cold front at the leading edge and negative (downward) vertical velocities over the lake.

The overriding atmosphere has an influence that is inversely proportional to the stability. Only during cool periods inland, when near neutral stabilities prevail on the lake, does the transfer of momentum and heat become efficient. This condition during the spring months is the exception (see Figure 32). Warm winds from the land create vigorous mixing at the upwind shore as they are uplifted over the cold lake air creating the characteristic inversion. Richardson numbers in excess of +0.5 are common within the cool but thermally stable lake anticyclone - demonstrating the minimized interaction of the gradient wind and the lake anticyclone over the central lake waters. Turbulent mixing, while prevalent over land, is nearly absent in the lake air-blanket where a much less turbulent flow persists.

The Great Lakes receive nearly all their heat, during the spring warming period, from insolation (Rodgers and Anderson 1961). Shallow nearshore regions receive additional heat from the atmosphere, where

the inversion is weaker and air-water interactions are possible, and river runoff concentrates. When the dynamic height of nearshore water becomes sufficiently unstable with the denser mid-lake waters through 1) excessive heat buildup around the lake-edge, and 2) the effective wind stress (transfer of energy through the inversion when the air temperature is only slightly above that of the water) the "thermal bar" disappears and the epilimnion develops across the entire lake.

The buoyant epilimnion warms more rapidly and reduces the air-water stability during summer. With this tendency toward more neutral air-water stability, the blanket of cold air is no longer produced, and the gradient wind gradually becomes more directly coupled with the water. Toward mid- to late-summer these more neutral conditions become increasingly frequent. The low level Richardson numbers reduce to values approaching zero, K_H approaches K_M, and the period of direct energy transfer begins. This transfer reaches a maximum as instability ($T_a << T_w$) typifies the fall and winter.

A schematic illustration of the lake effect when $T_a >> T_w$ is shown in Figure 47. The prevailing gradient wind is presented as normal-to-shore. This flow was chosen because it demonstrates the extreme interactions at the shoreline: maximum at the upwind shore, and minimum at the downwind shore. The slope of the surface of the water has been vertically exaggerated as in Figure 31. The lake has warmed substantially to the stage that the thermocline is in its developing stages. While subsidence characterizes the vertical velocity over the lake it is greatest at the upwind shore. The wind stress, while considerably reduced, will generally pile up more epilimnetic water along the downwind shore than at the upwind shore. Under strong inversion conditions heat

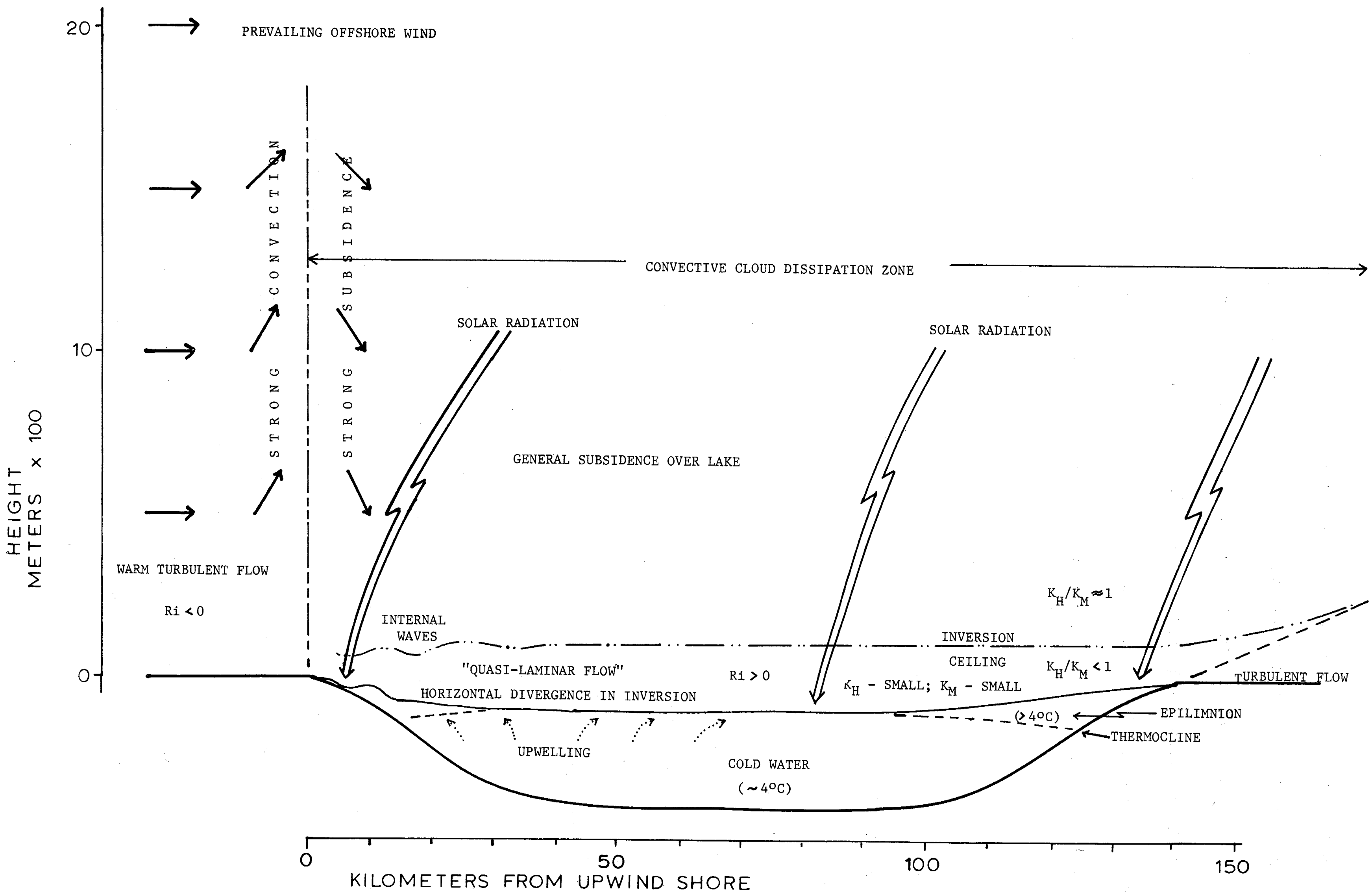

FIG. 47. A vertical cross section of a physical model of the environment of a large temperate lake during spring under normal-to-shore winds.

transfer is quite small as the low level air flow approaches a "quasi-laminar" state. K_H will be effective only in the limited regions of warmer water near the upwind shore. In these regions, where the inversion has not yet become intense, the transfer of heat and momentum into the water, plus river runoff, slowly warms the water and assists in developing the thermocline.

At the downwind shore the lake air becomes modified as its trajectory carries cool air over the heated land. Observations (Moroz 1967) show that the inversion, while being rapidly reduced landward, increases the height of its ceiling. Buoyant forces, developing in the modifying lake air, gradually destroy the inverted thermal characteristics of the lake air. When the inversion is finally dispersed, growth of convective cumulus will again become possible. The extent of this cumulus suppression zone inland must be a function of not only the air-water temperature difference and insolation, but also of the gradient wind intensity. While the vertical effect of the lake at the downwind shore is limited, the horizontal extent of lake effects inland may be quite large. Recent satellite cloud pictures have shown downwind lake effects that suppress cumulus activity occasionally as far as several tens of kilometers (Parmenter 1967).

CONCLUSIONS

While the Great Lakes have been used previously for observational studies of the stable air-water ($T_{air} > T_{water}$) interaction, the present incorporation of Estoque's numerical model for the sea breeze has shown that for most studies a body of water that has a horizontal extent of 100 kilometers or more can be classified as an "ocean" when considering these effects. The Great Lakes provide separate lake effect atmospheric circulations over each lake when $T_{air} > T_{water}$. These effects may extend several tens of kilometers inland from shore but more customarily are contained within the first 10 kilometers.

A direct consequence of the stable period is the absence of convective cloudiness (low level) over the lake. Richards and Loewen (1965) have shown solar radiation to be as much as 20% higher over the Lakes than over land during this period. If cumulus clouds are advected by the wind over the cold lake, the subsidence which characterizes the center of the lake anticyclone rapidly dissipates these clouds. The sun continues to supply thermal energy into the surface waters but low level wind divergence and mid-lake upwelling prevent any immediate major concentration of this heat in the surface waters in the central region of the lake. The heat energy is stored around the edge of the lake and a strong thermal gradient, that decreases toward mid-lake, develops. This gradient provides the major tendency for current circulation of the spring and early summer - cyclonic.

Air-water interaction during the stable months may inhibit or enhance nearshore geostrophic water transport depending on the prevailing wind direction over the basic cyclonic thermally induced circulation. The

prevailing wind stress for the Great Lakes region during spring and summer is from the south and west. Cyclonic thermal lake circulation under these conditions will be best developed along the southern and eastern shores of the lakes, while weaker more diffuse currents and smaller eddies will characterize the other shores. As the epilimnion develops and reduces the surface temperature gradient the large basin gyre breaks down into smaller scale circulations which have been recently investigated by Noble (1967).

APPENDIX

Daily satellite pictures
Daily mesoscale weather analyses
Daily vorticity analyses
May and June weekly water temperatures for 1966

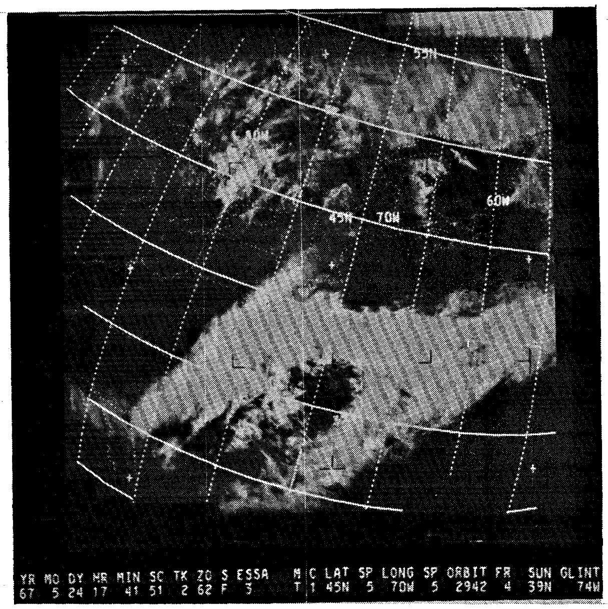

FIG. A1. ESSA 3 cloud picture of the Great Lakes - 24 May 1967: 1241:51E.

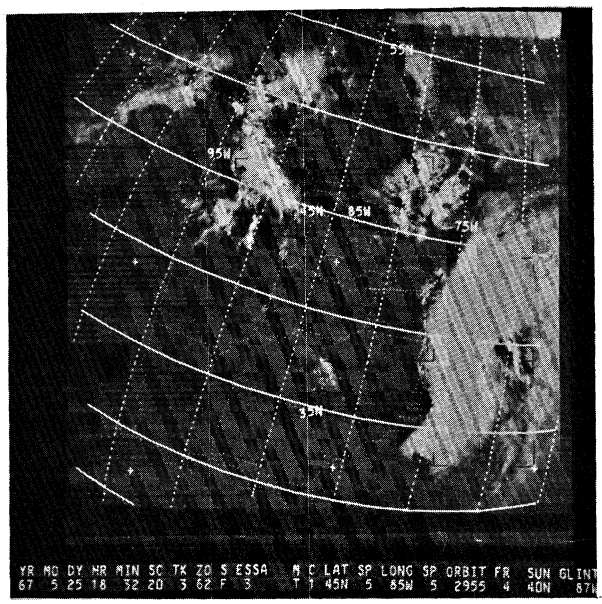

FIG. A2. ESSA 3 cloud picture of the Great Lakes - 25 May 1967: 1332:20E.

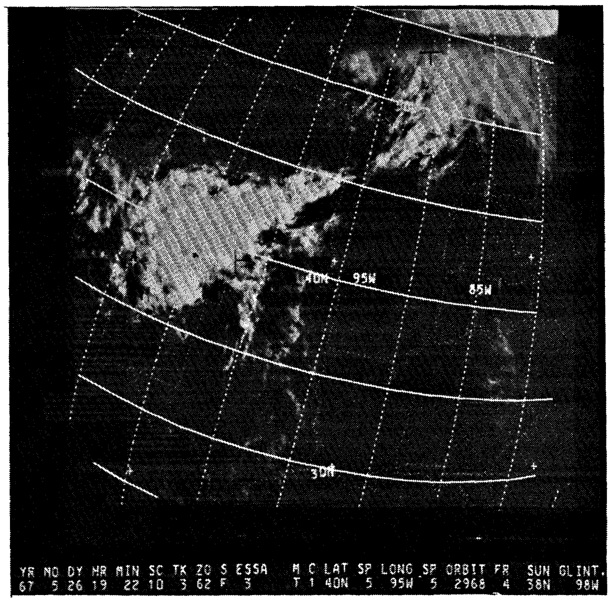

FIG. A3. ESSA 3 cloud picture of the Great Lakes - 26 May 1967: 1422:10E.

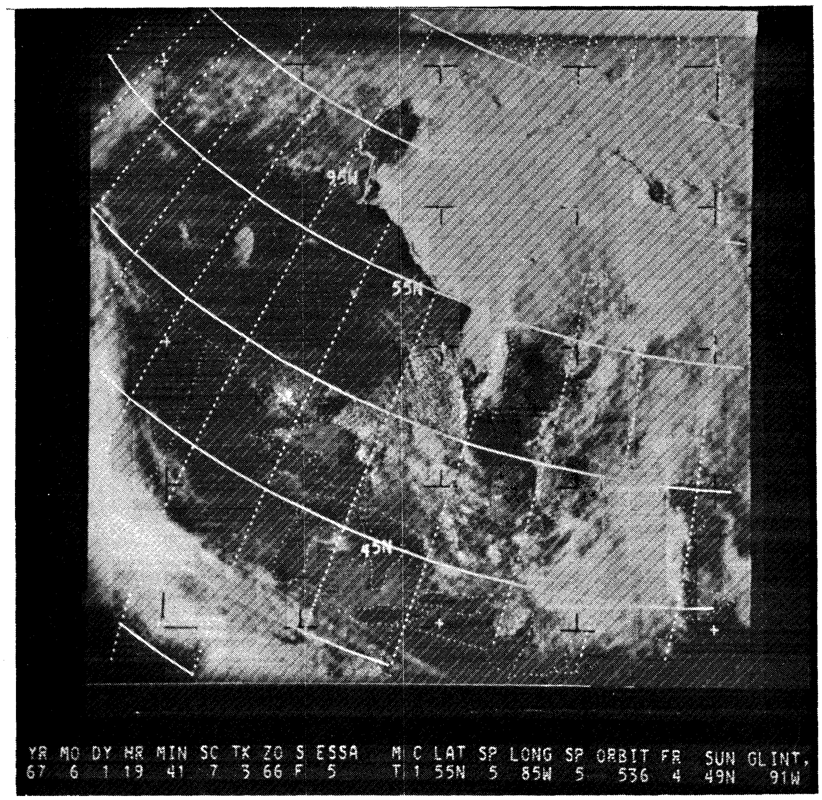

FIG. A4. ESSA 5 cloud picture of the Great Lakes - 1 June 1967: 1441:07E.

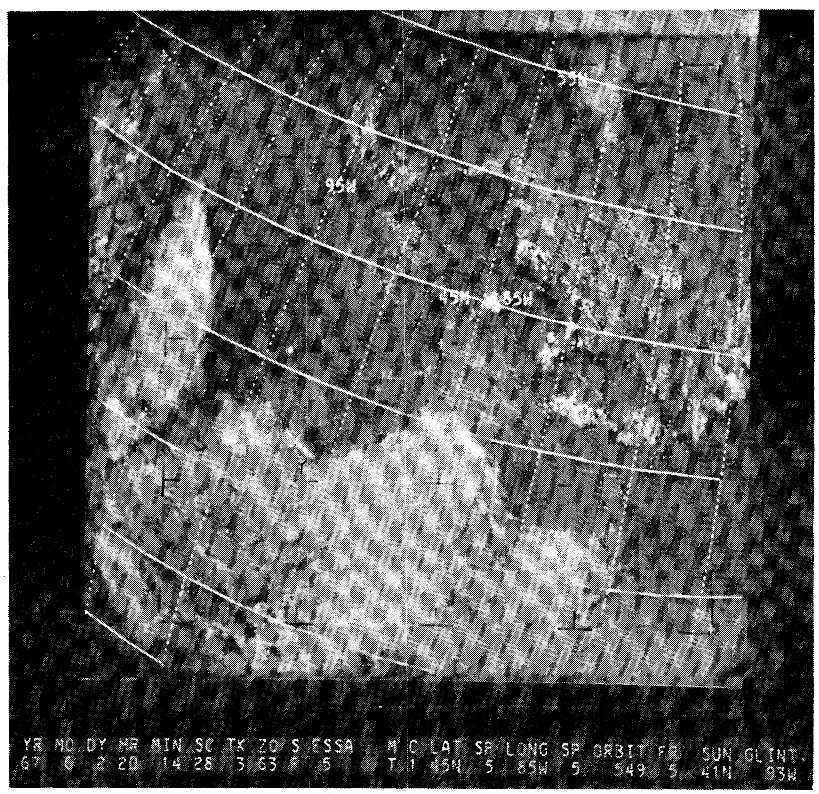

FIG. A5. ESSA 5 cloud picture of the Great Lakes - 2 June 1967: 1514:28E.

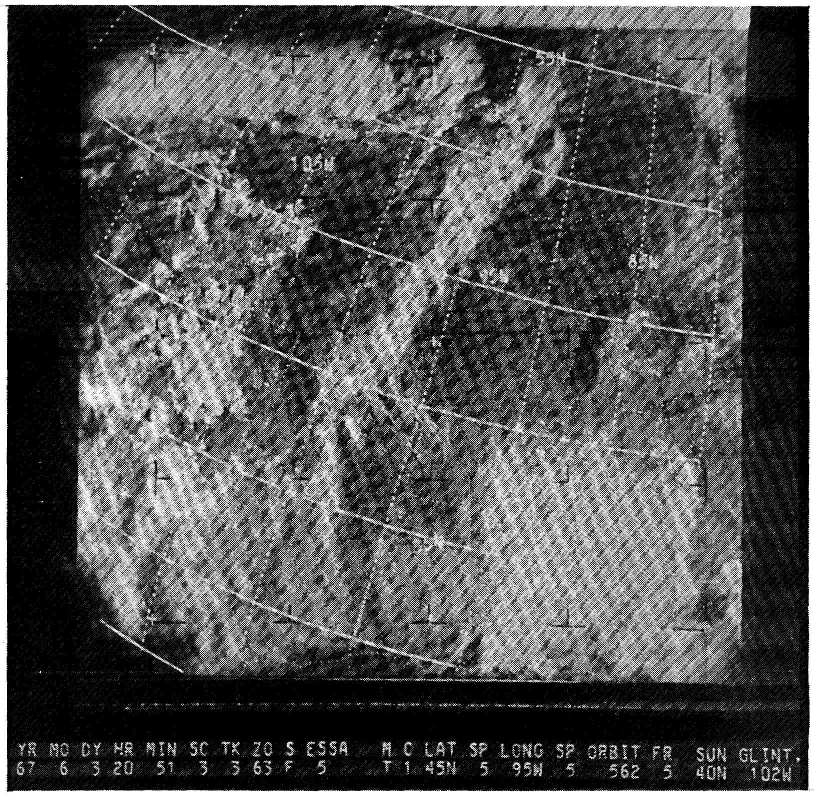

FIG. A6. ESSA 5 cloud picture of the Great Lakes - 3 June 1967: 1551:03E.

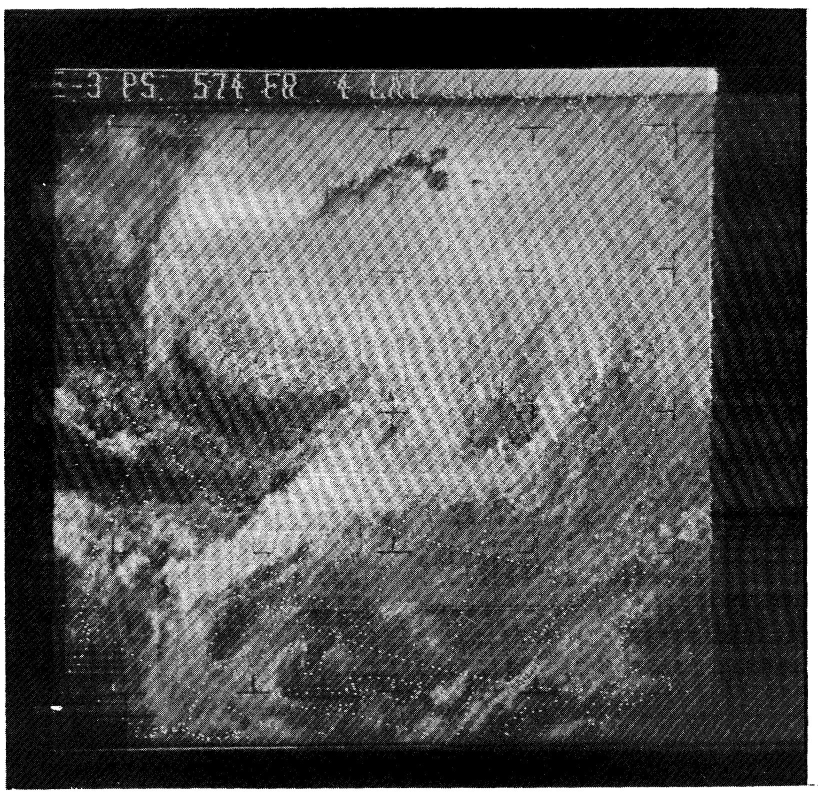

FIG. A7. ESSA 5 cloud picture of the Great Lakes - 4 June 1967: 1438:11E.

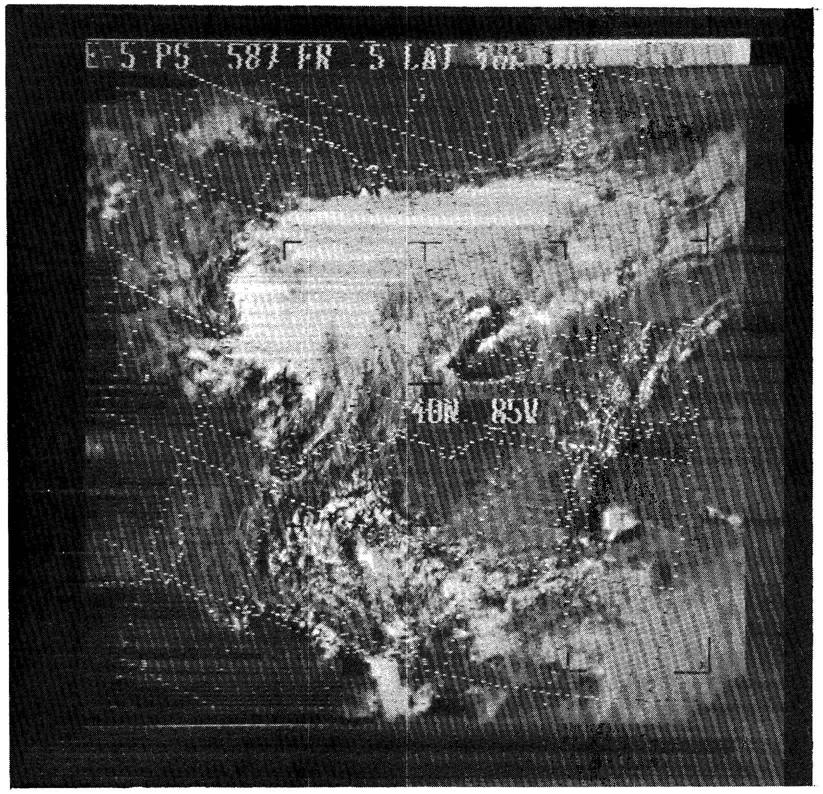

FIG. A8. ESSA 5 cloud picture of the Great Lakes - 5 June 1967: 1510:29E.

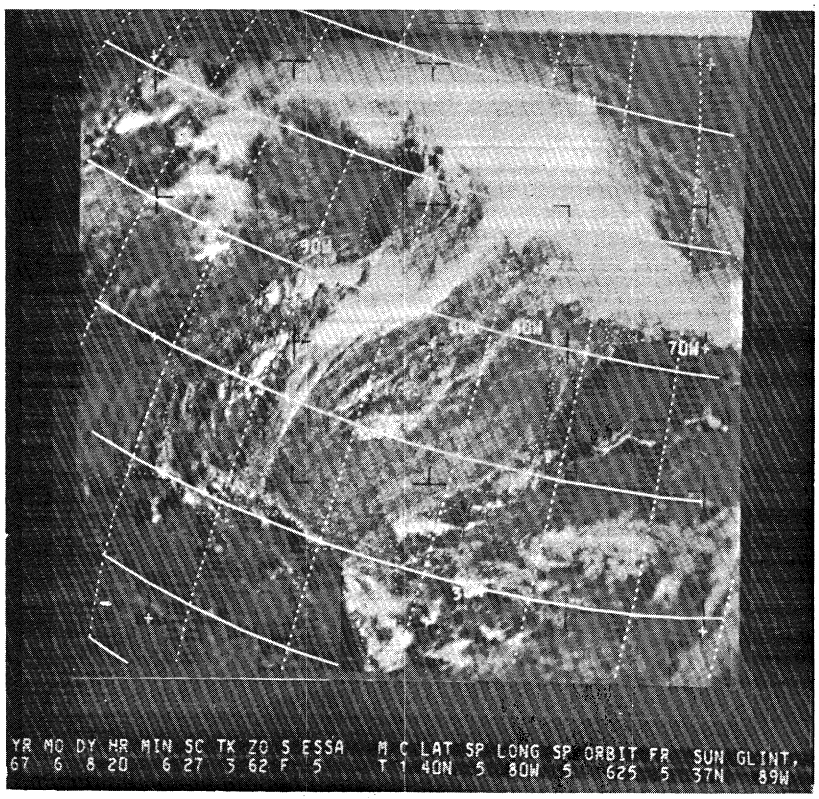

FIG. A9. ESSA 5 cloud picture of the Great Lakes - 8 June 1967: 1506:27E.

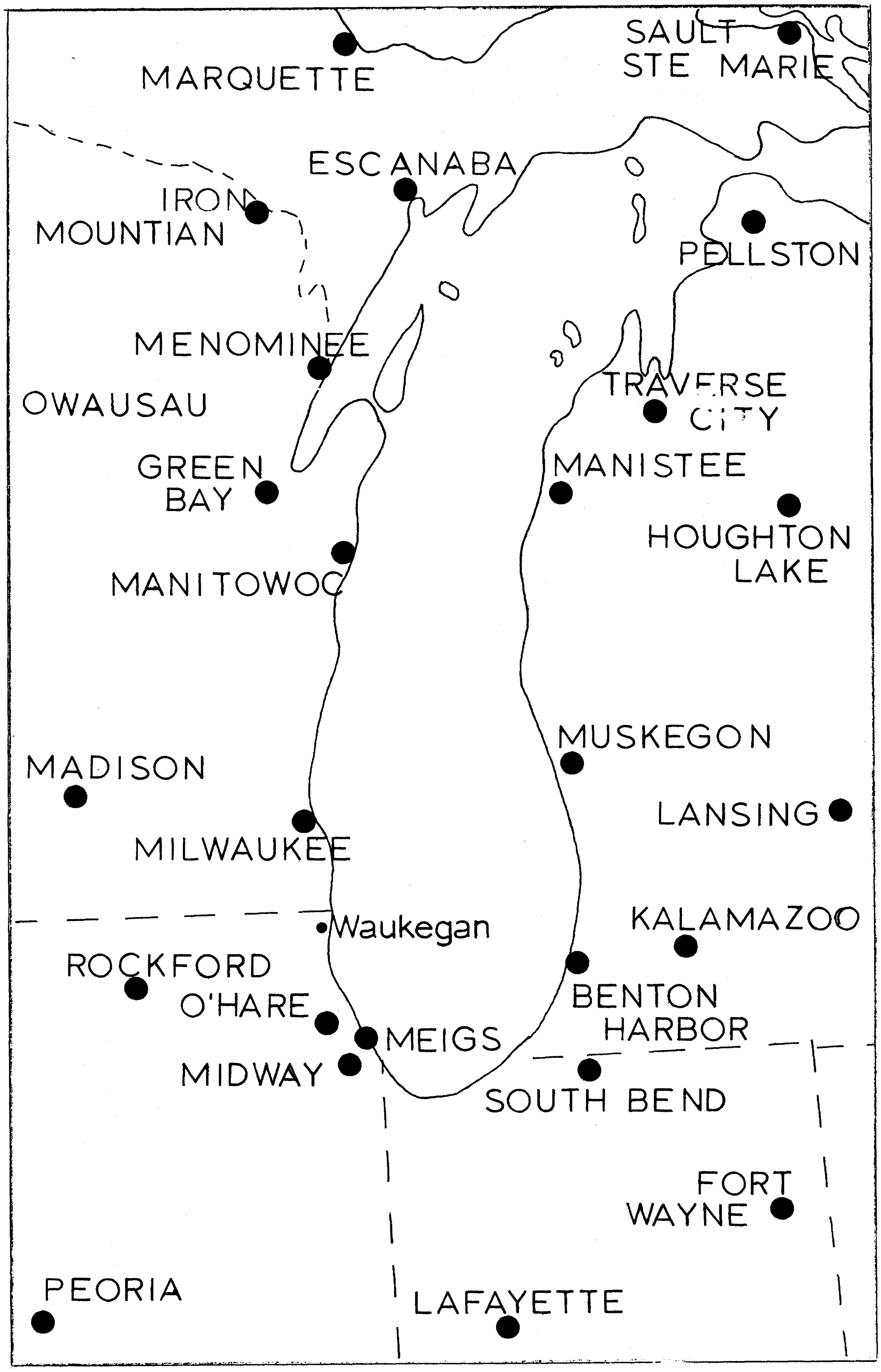

FIG. A10. Mesoscale stations used in the following analyses. For additional stations used see text.

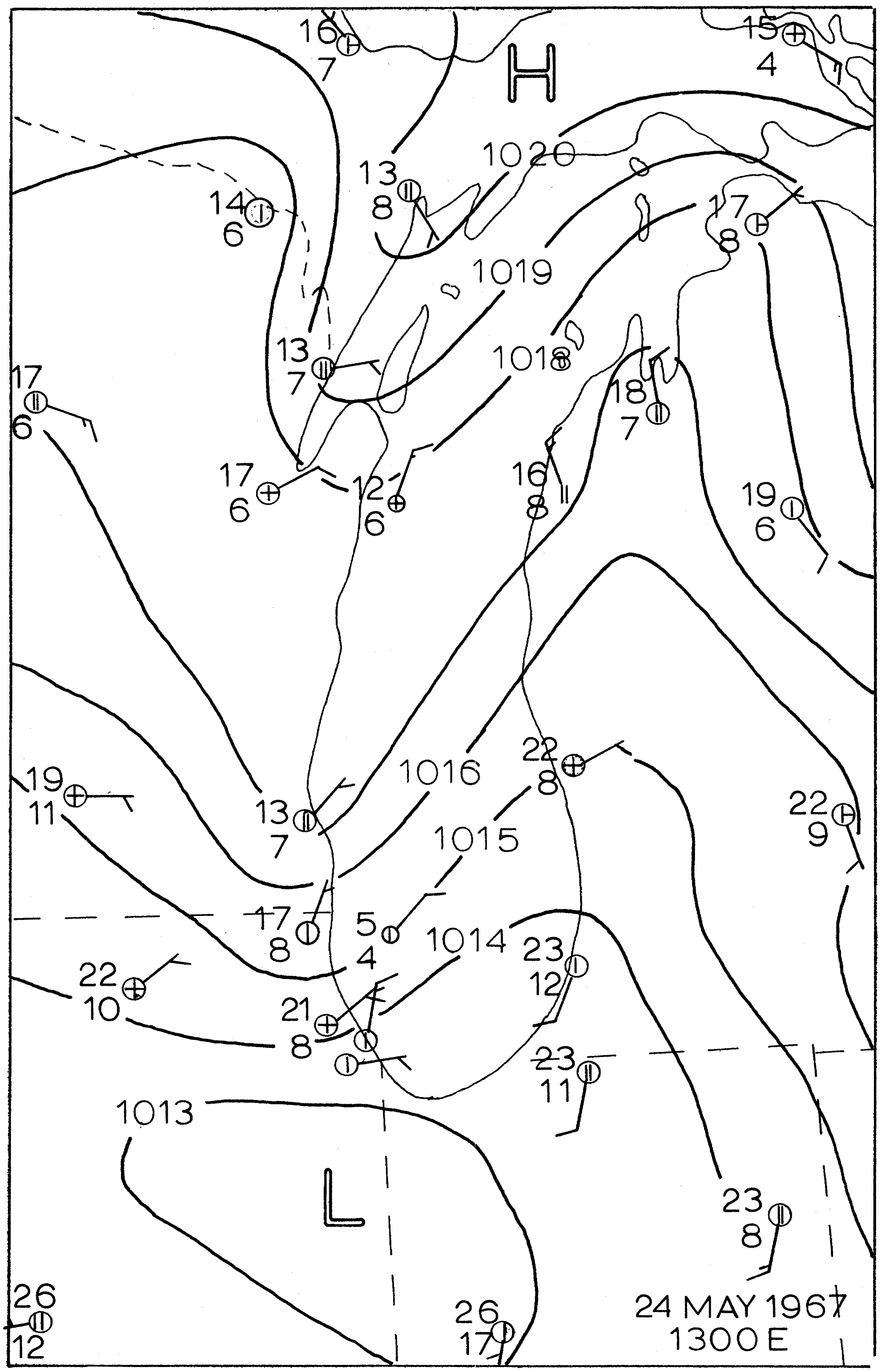

FIG. A11. Mesoscale weather - 24 May 1967 at 1300E. Isobars are in millibars. Air temperature (upper) and dew point temperature (lower) appear to the left of station - °C.

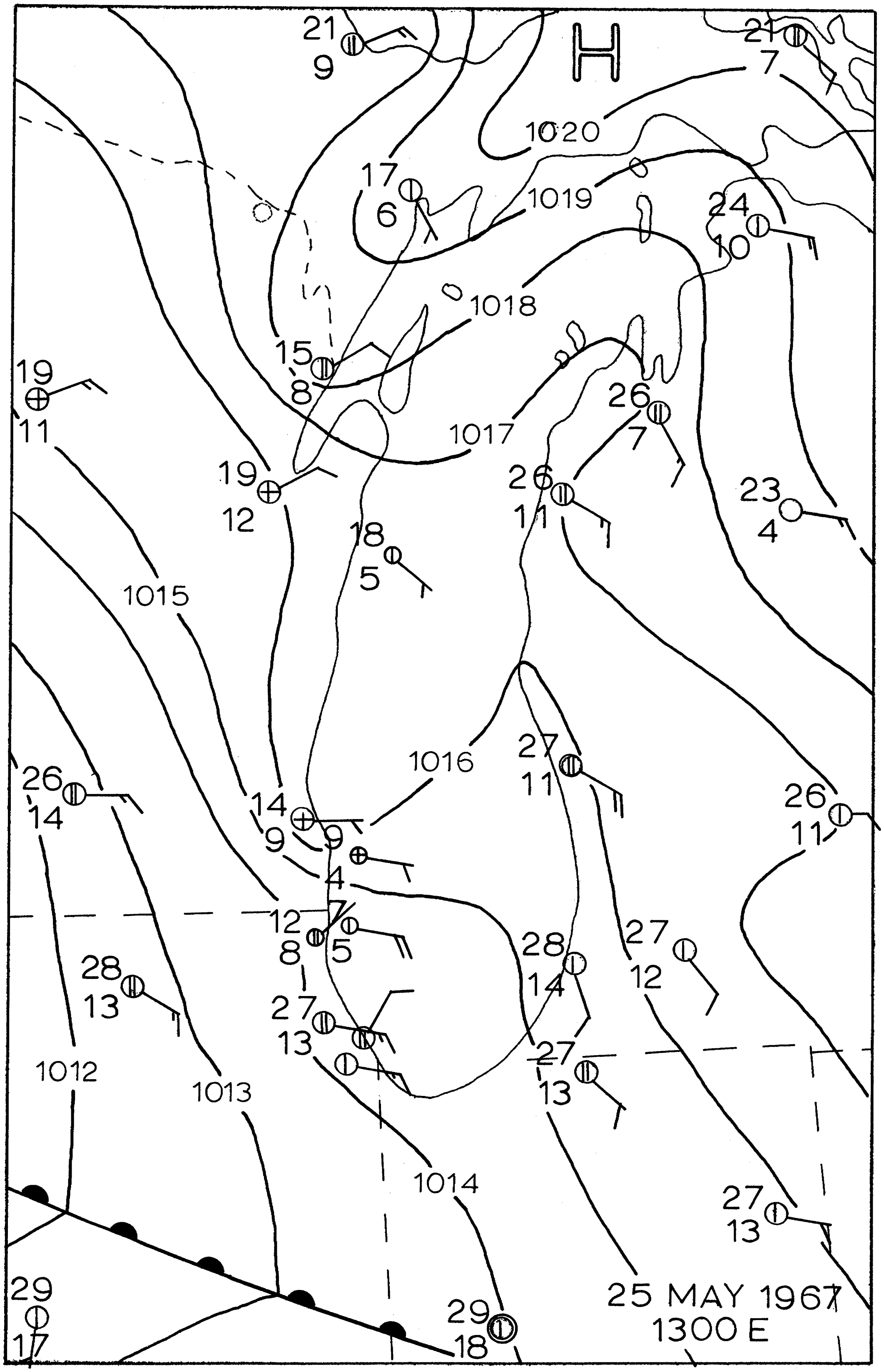

FIG. A12. Mesocsale weather - 25 May 1967 at 1300E. Details on Fig. A11.

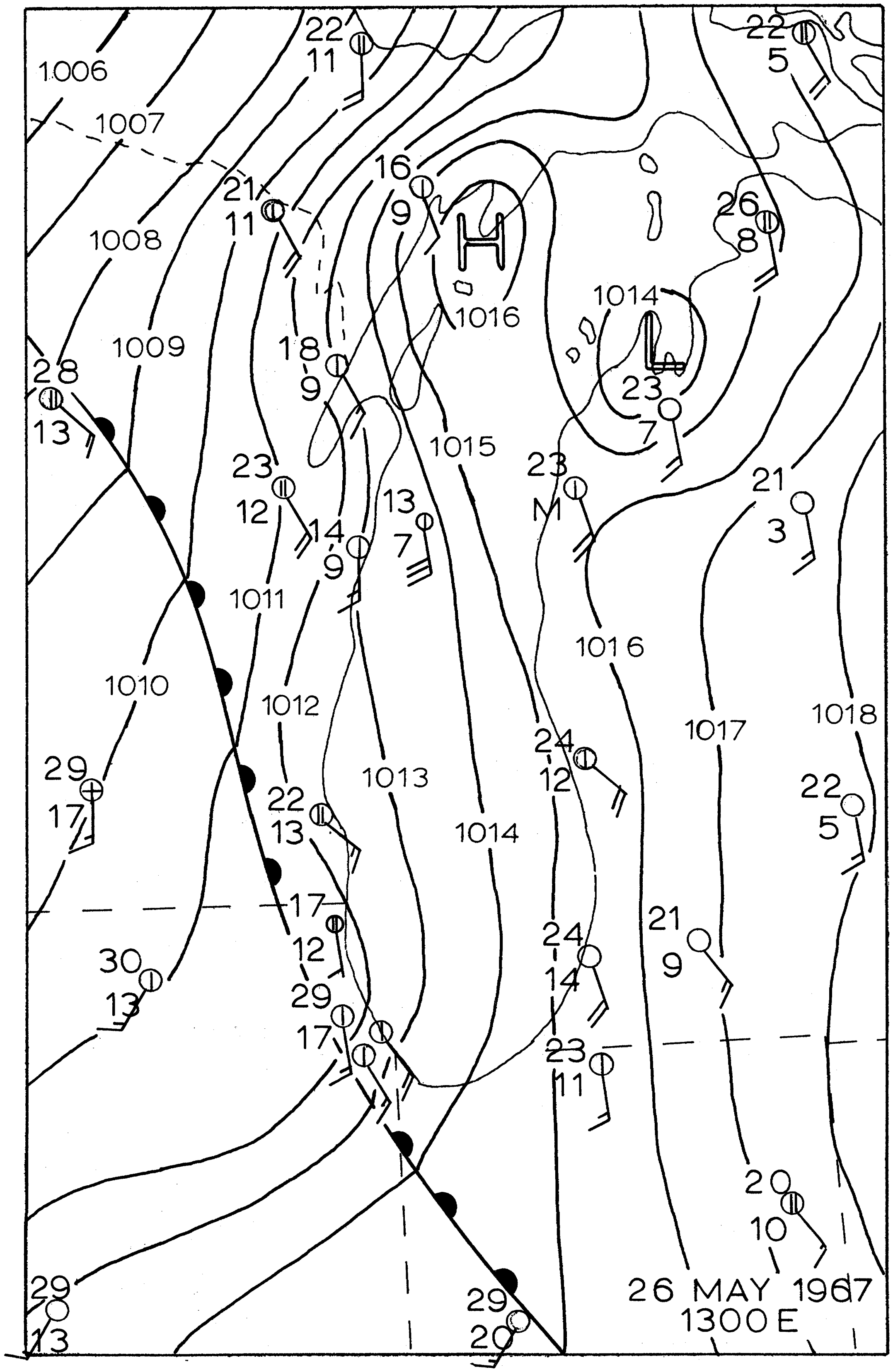

FIG. A13. Mesoscale weather - 26 May 1967 at 1300E. Details on Fig. A11.

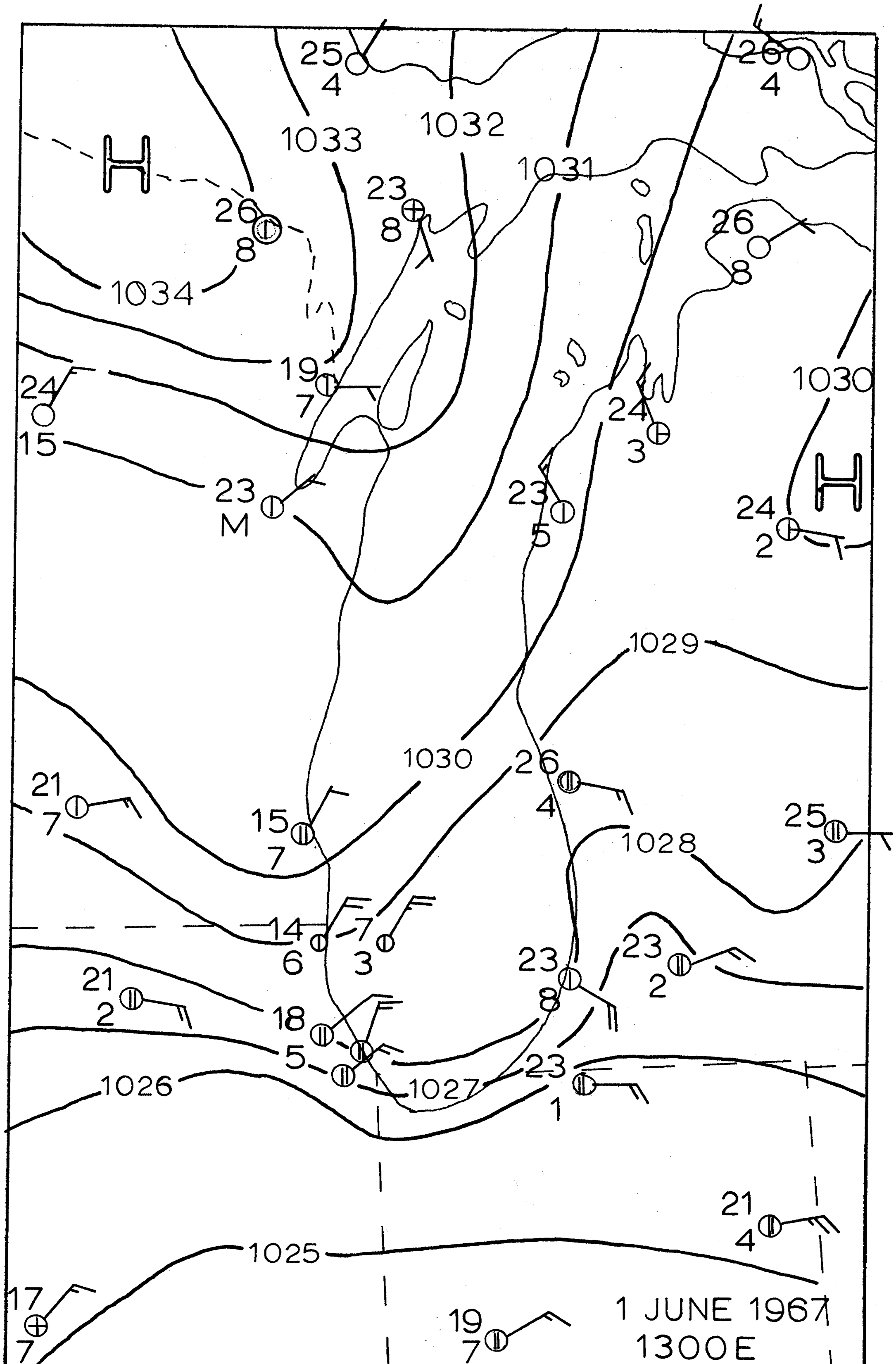

FIG. A14. Mesoscale weather - 1 June 1967 at 1300E. Details on Fig. A11.

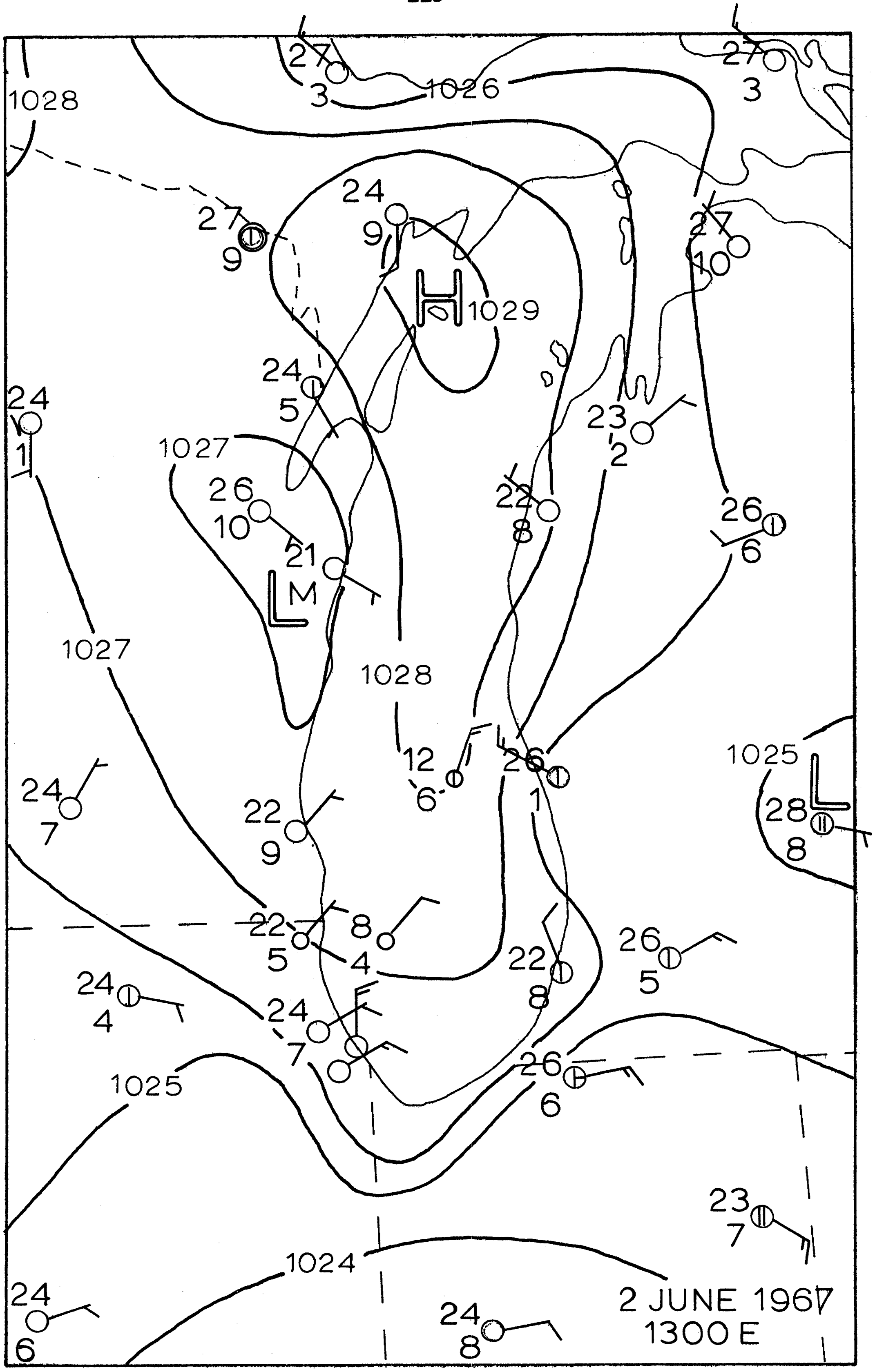

FIG. A15. Mesoscale weather - 2 June 1967 at 1300E. Details on Fig. A11.

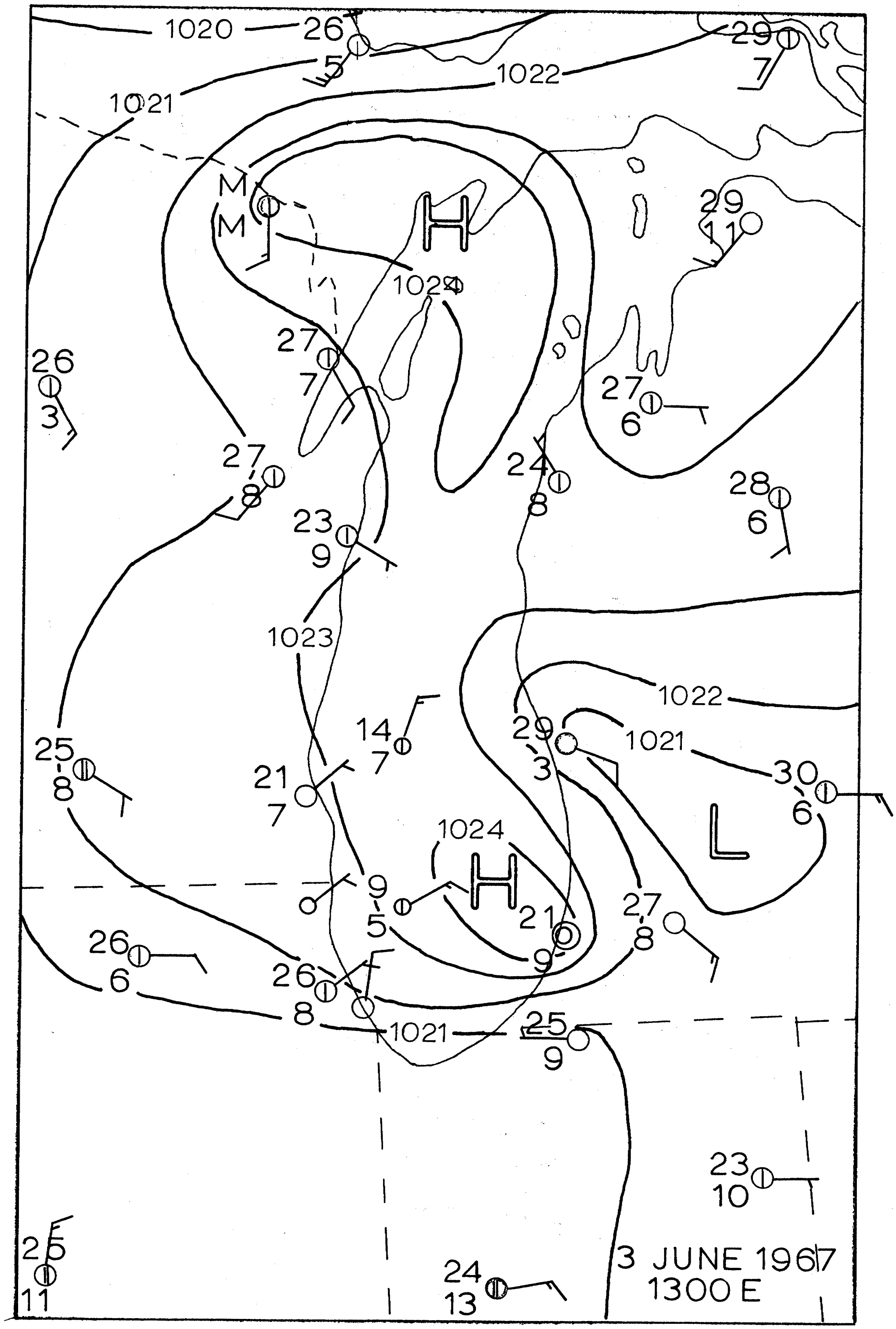

FIG. A16. Mesoscale weather - 3 June 1967 at 1300E. Details on Fig. All.

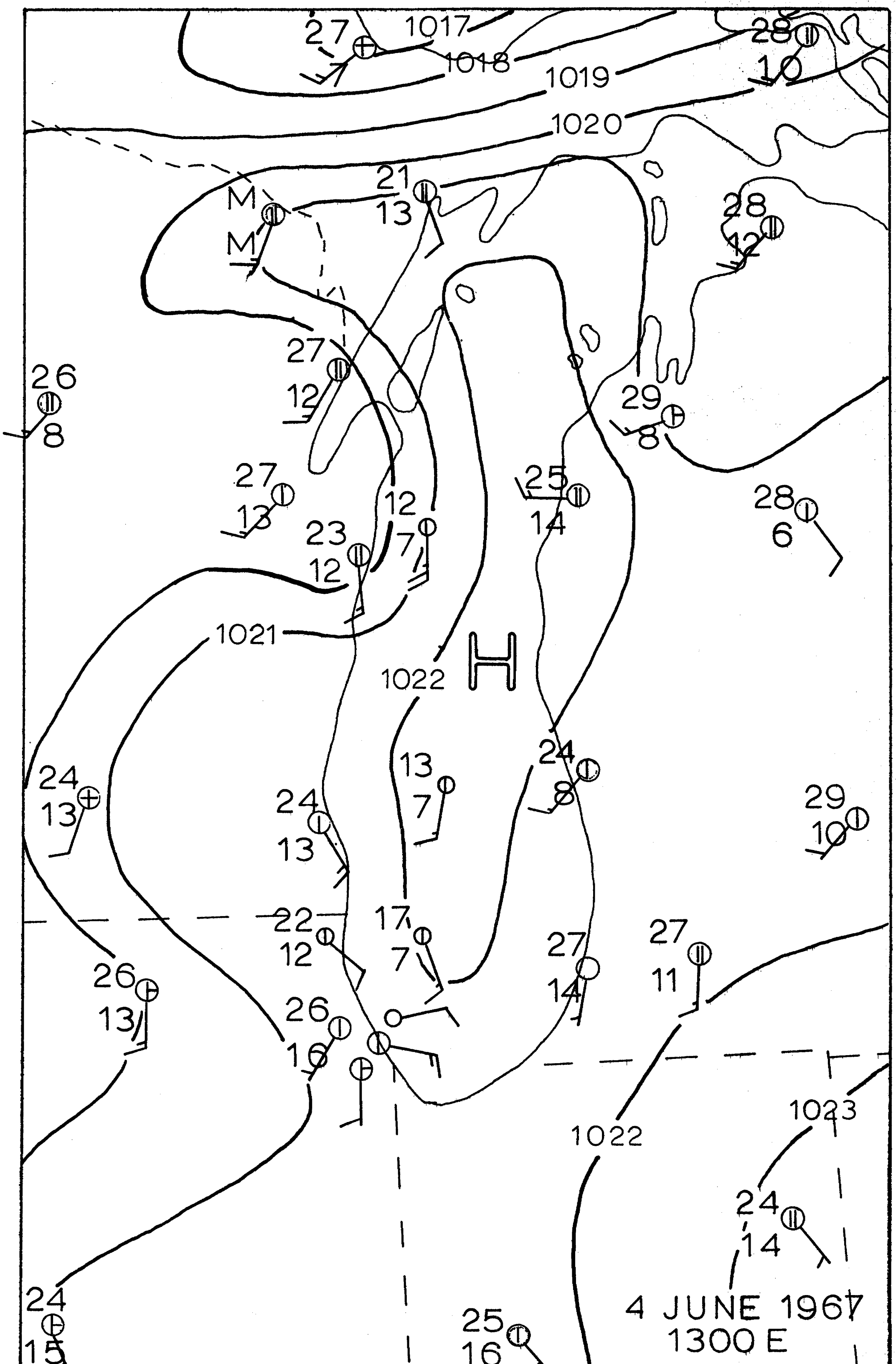

FIG. A17. Mesoscale weather - 4 June 1967 at 1300E. Details on Fig. A11.

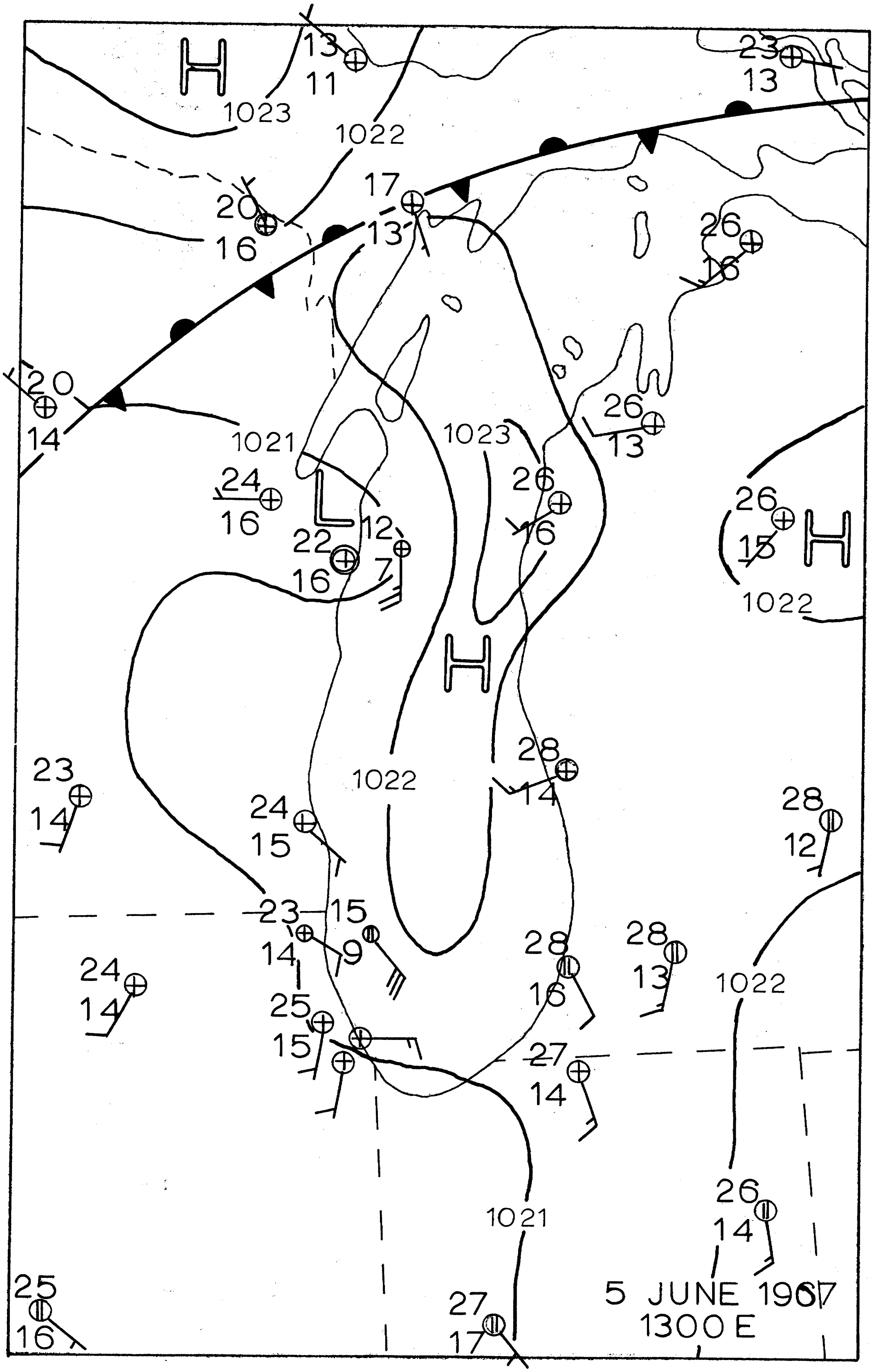

FIG. A18. Mesoscale weather - 5 June 1967 at 1300E. Details on Fig. All.

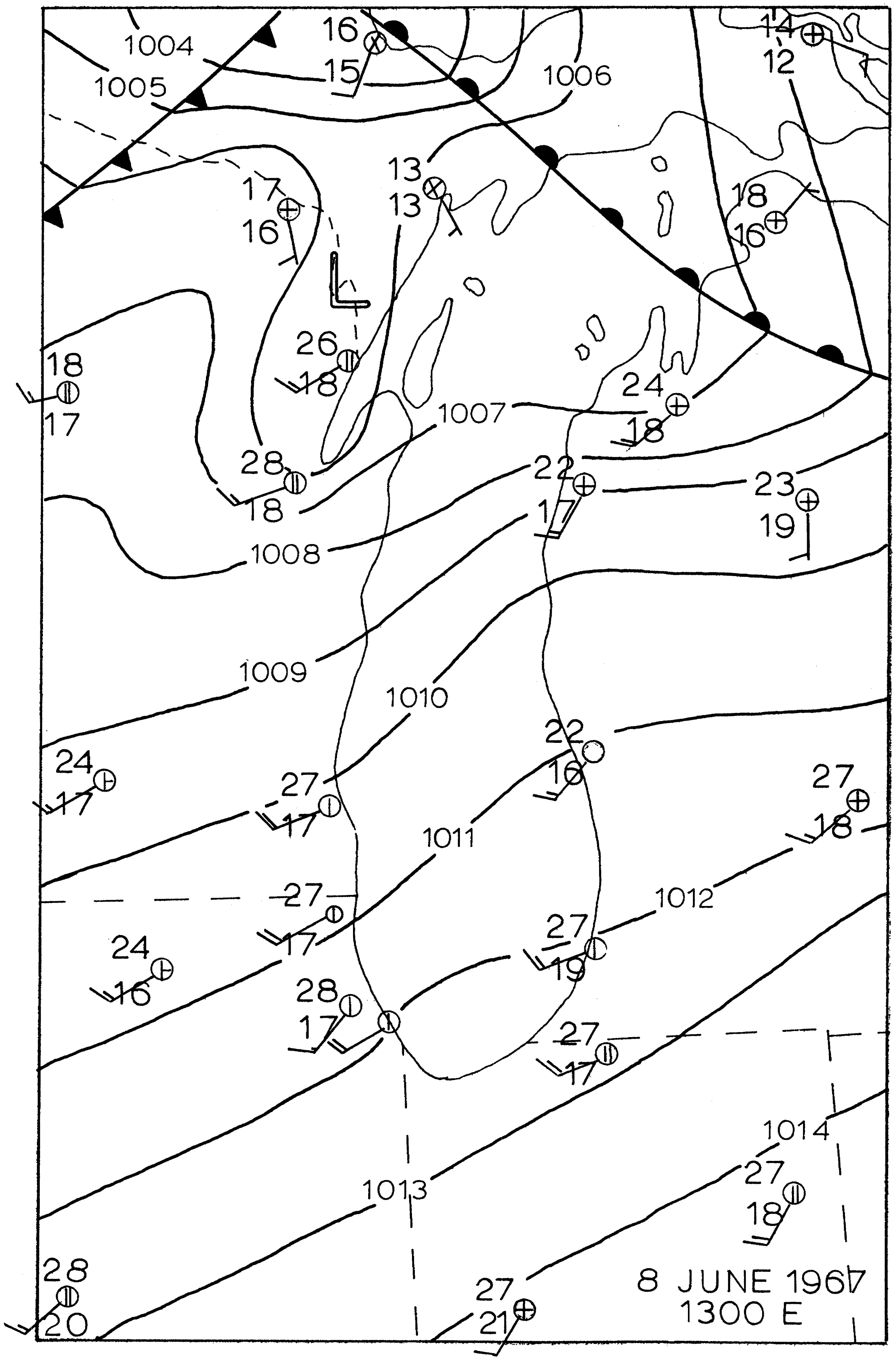

FIG. A19. Mesoscale weather - 8 June 1967 at 1300E. Details on Fig. A11.

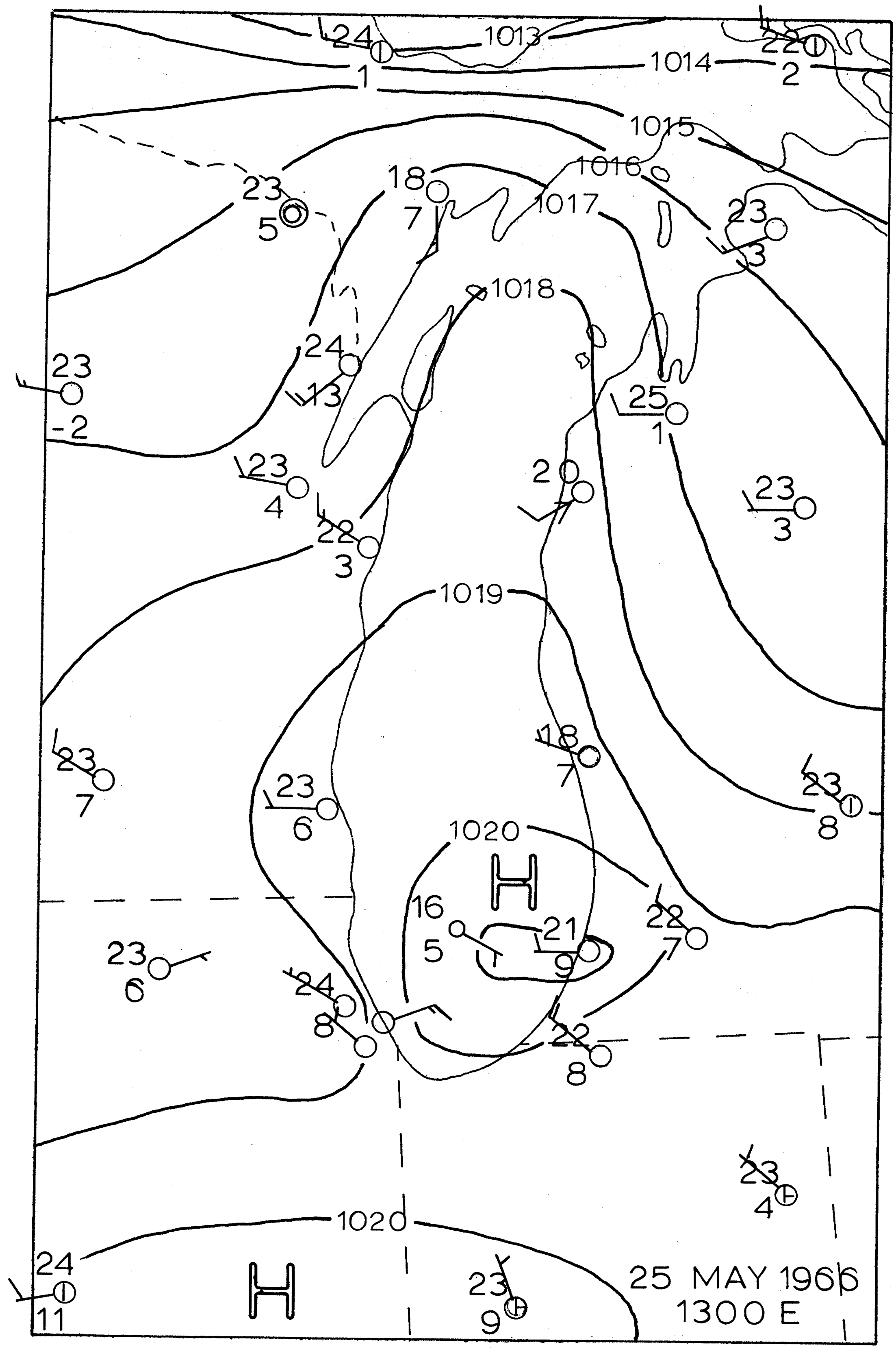

FIG. A20. Mesoscale weather - 25 May 1966 at 1300E. Details on Fig. A11.

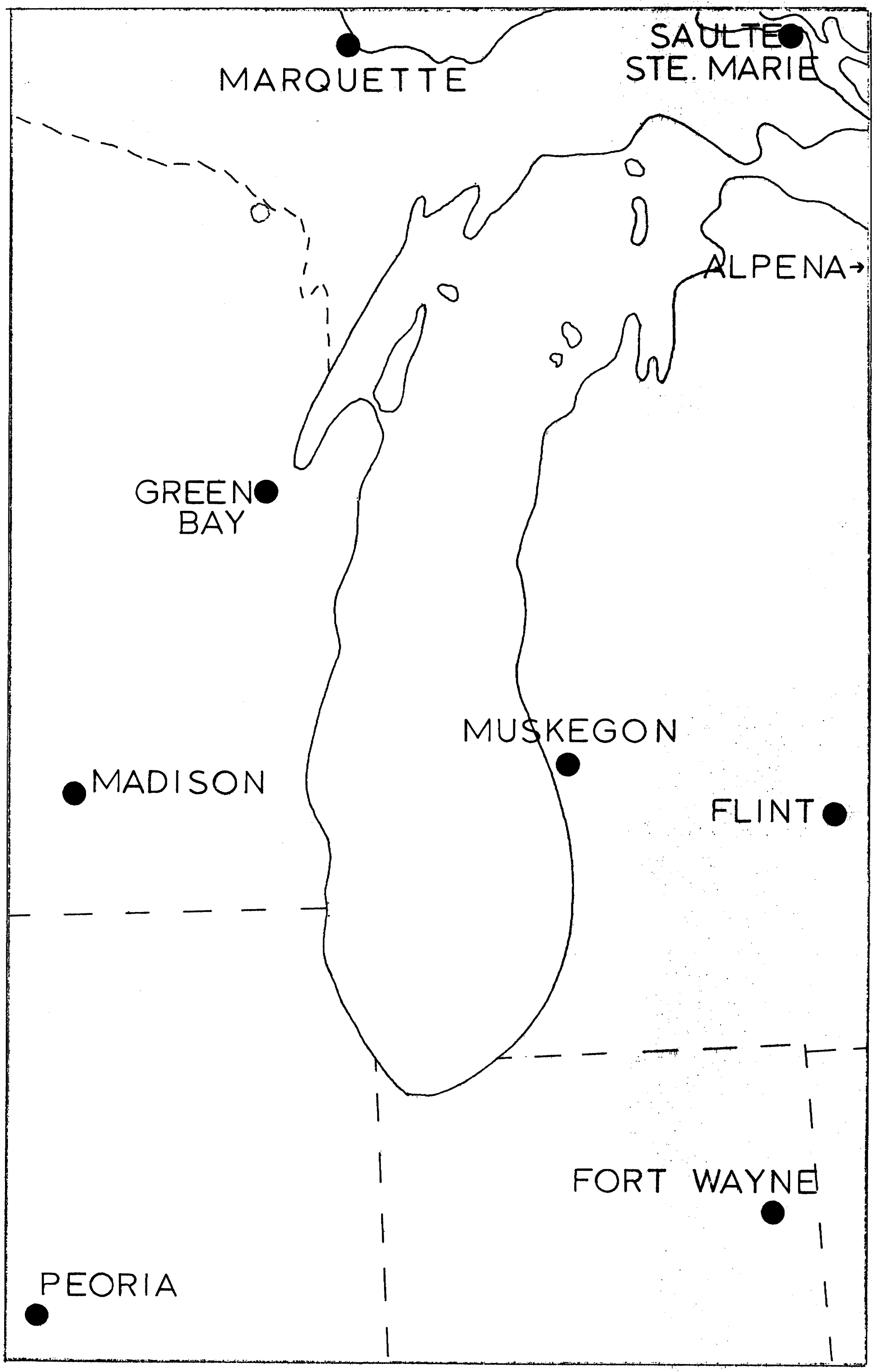

FIG. A21. 610 meter wind stations used in the following analyses.

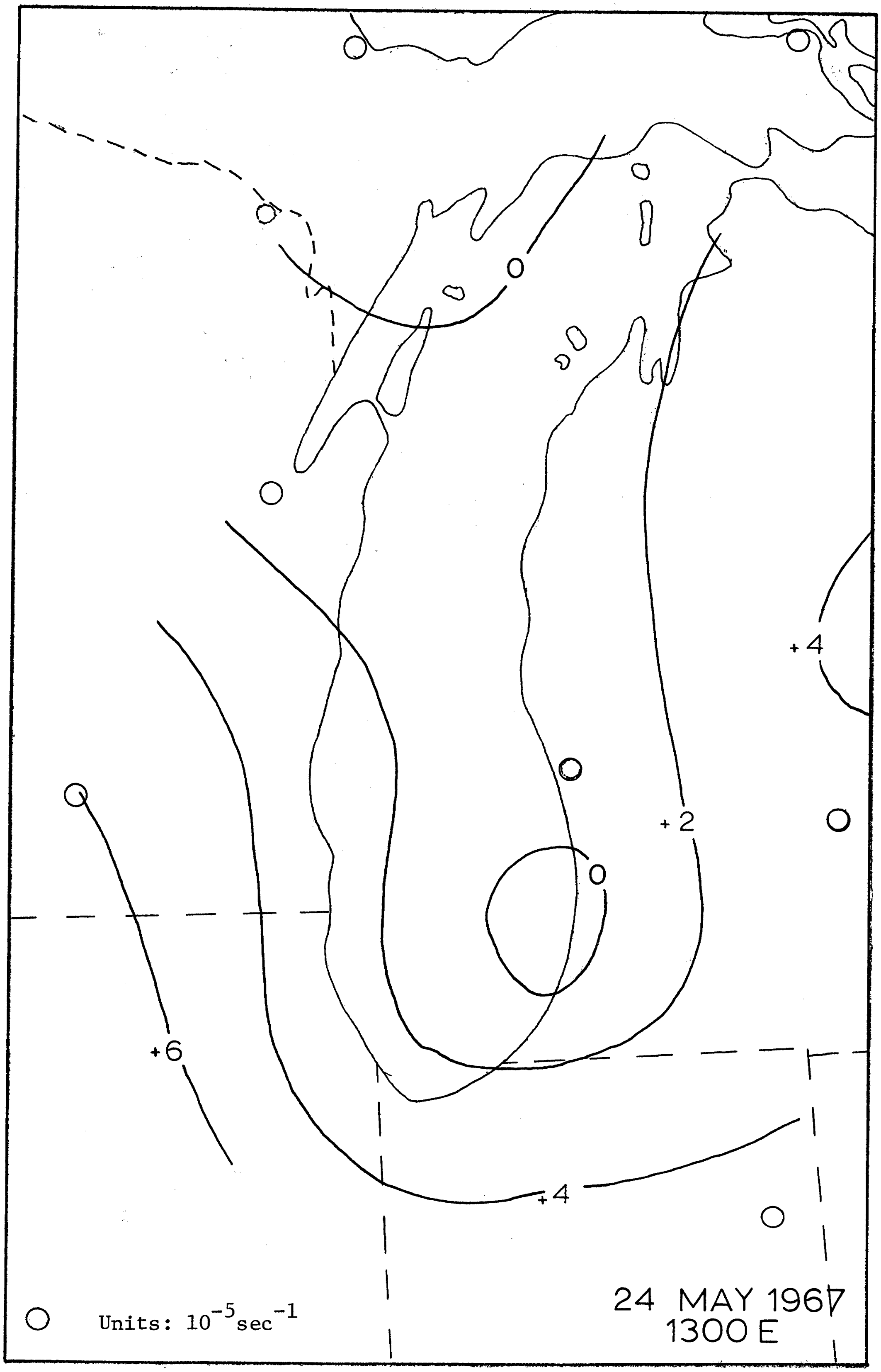

FIG. A22. 610 m absolute vorticity - 24 May 1967 at 1300E.

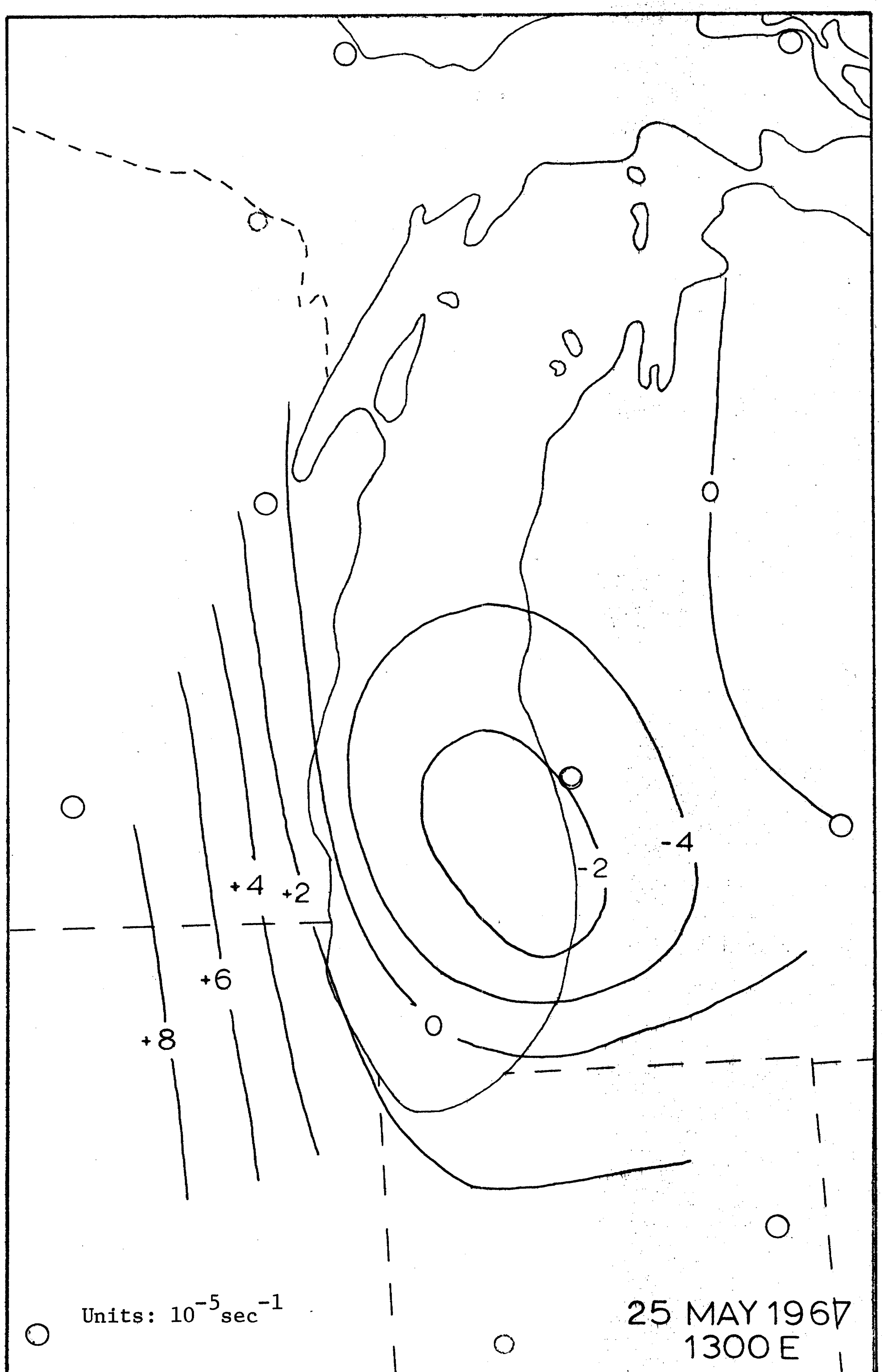

FIG. A23. 610 m absolute vorticity - 25 May 1967 at 1300E.

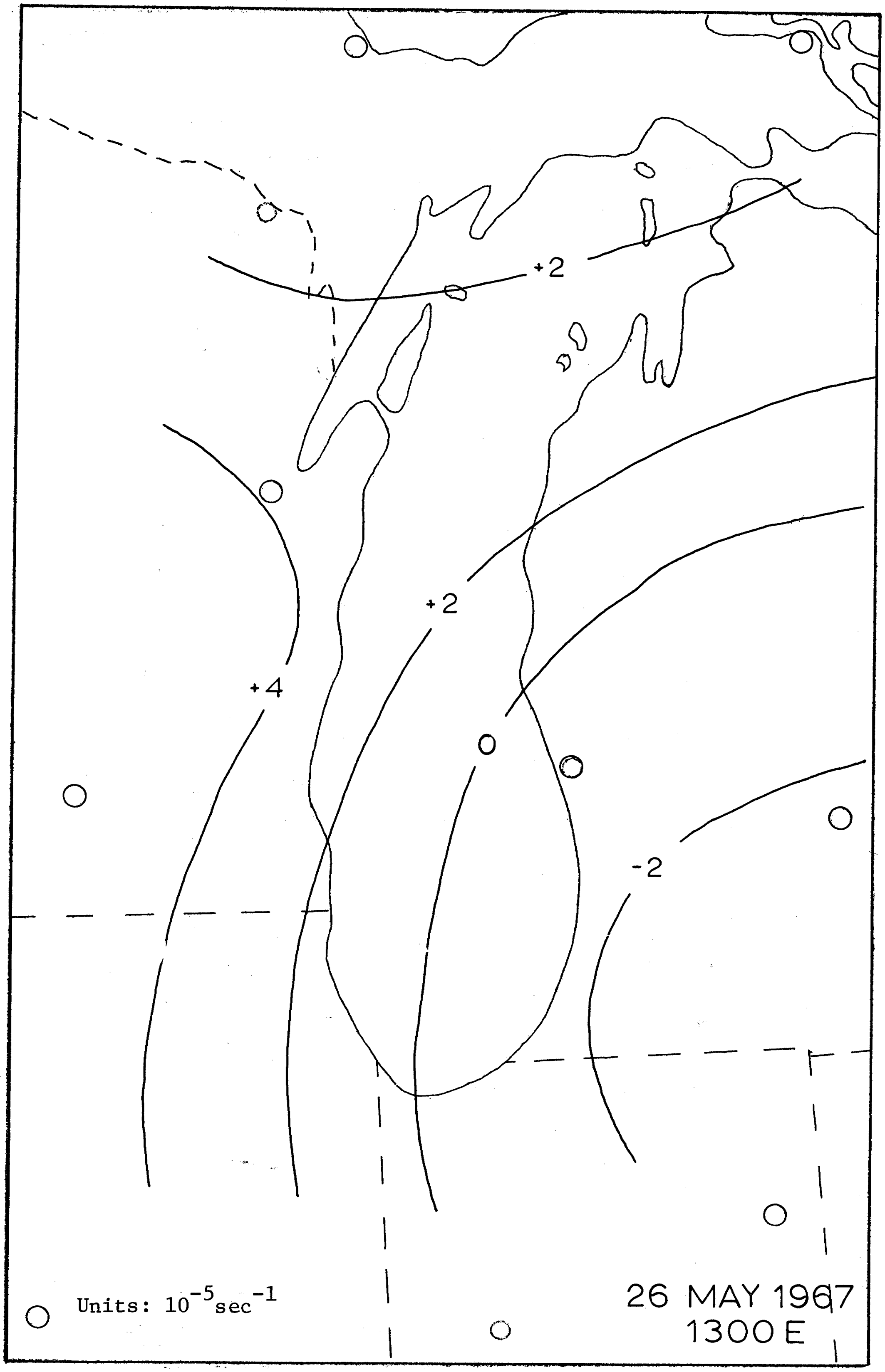

FIG. A24. 610 m absolute vorticity - 26 May 1967 at 1300E.

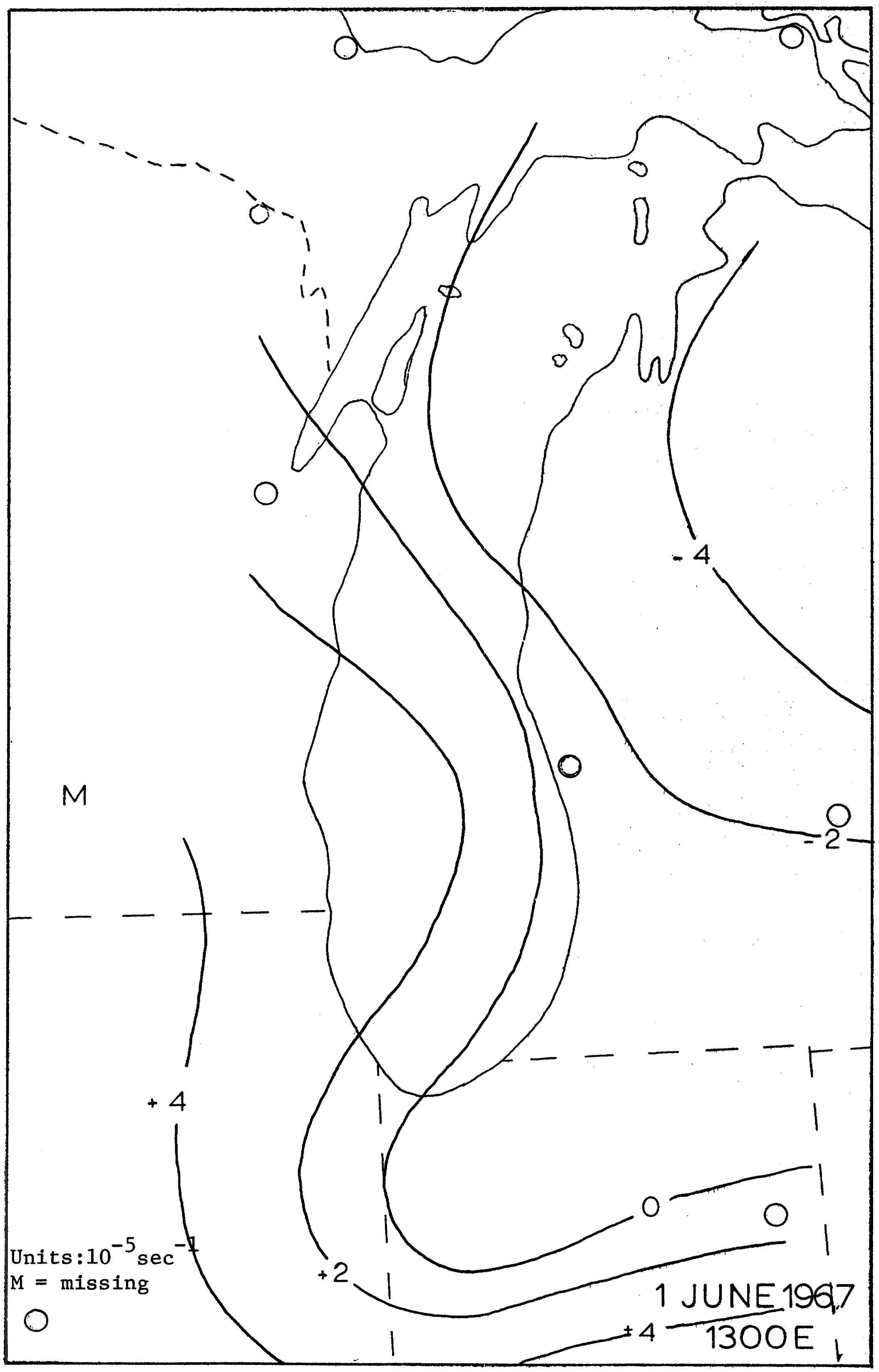

FIG. A25. 610 m absolute vorticity - 1 June 1967 at 1300E.

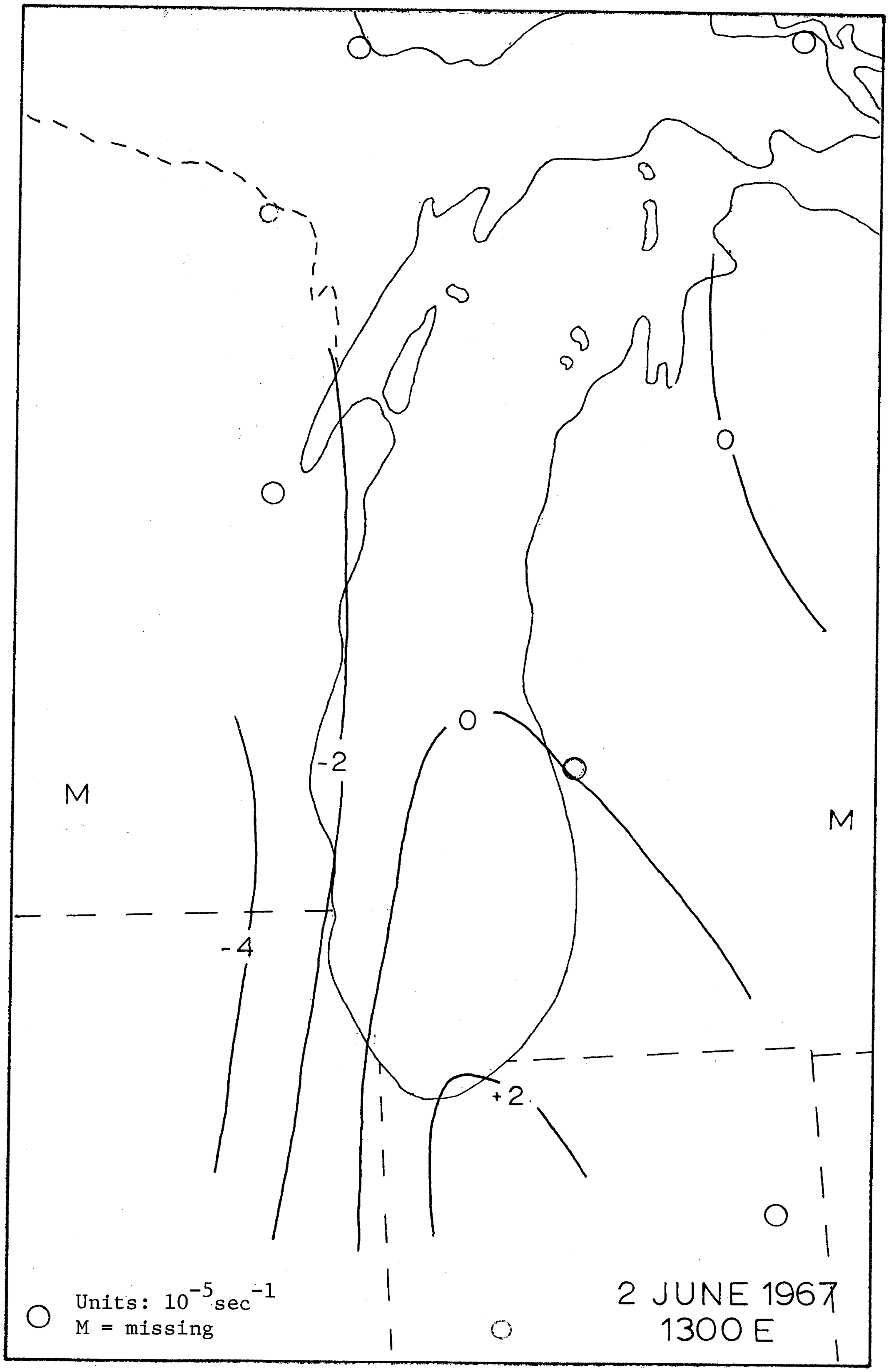

FIG. A26. 610 m absolute vorticity - 2 June 1967 at 1300E.

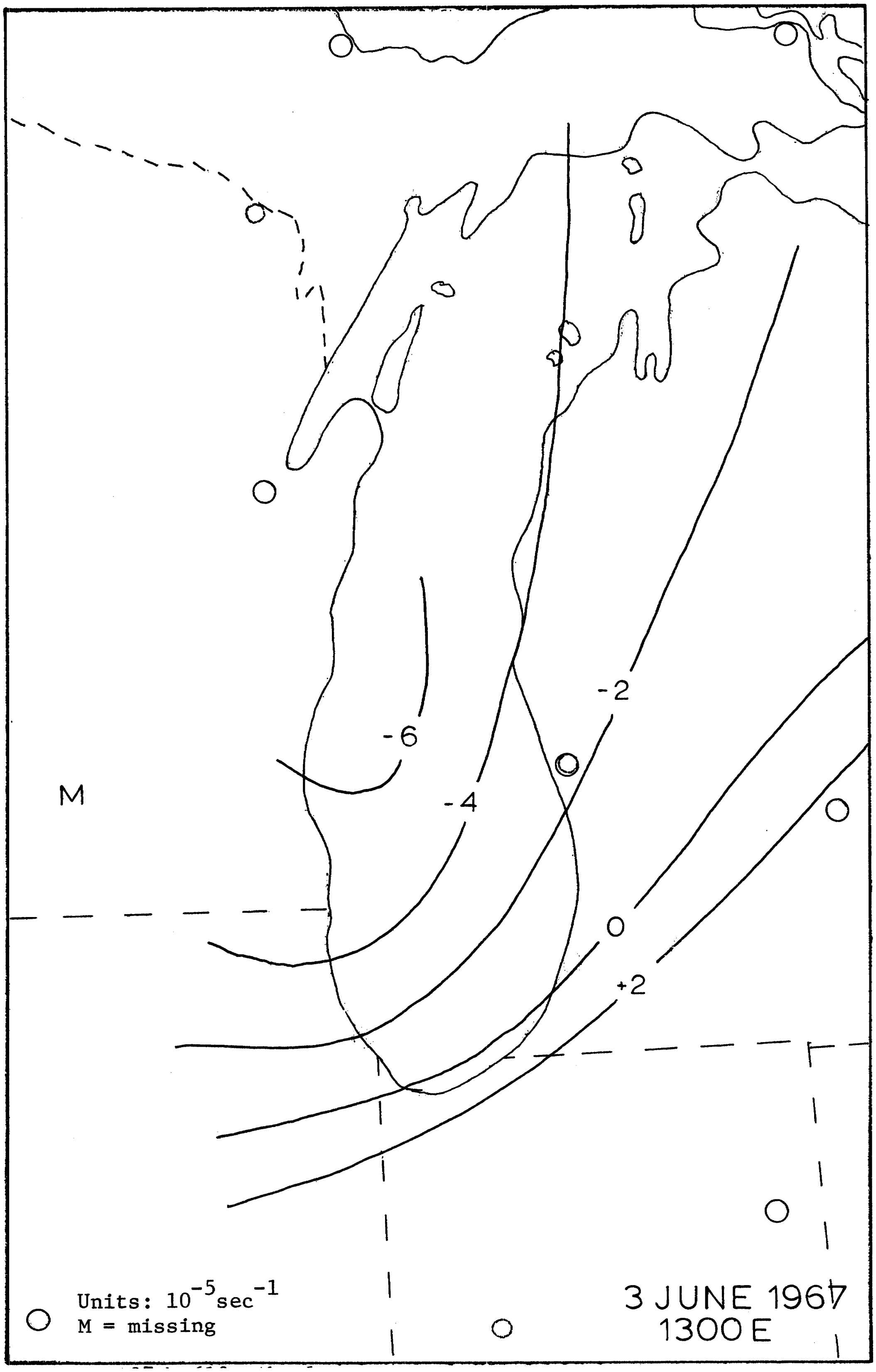

FIG. A27. 610 m absolute vorticity - 3 June 1967 at 1300E.

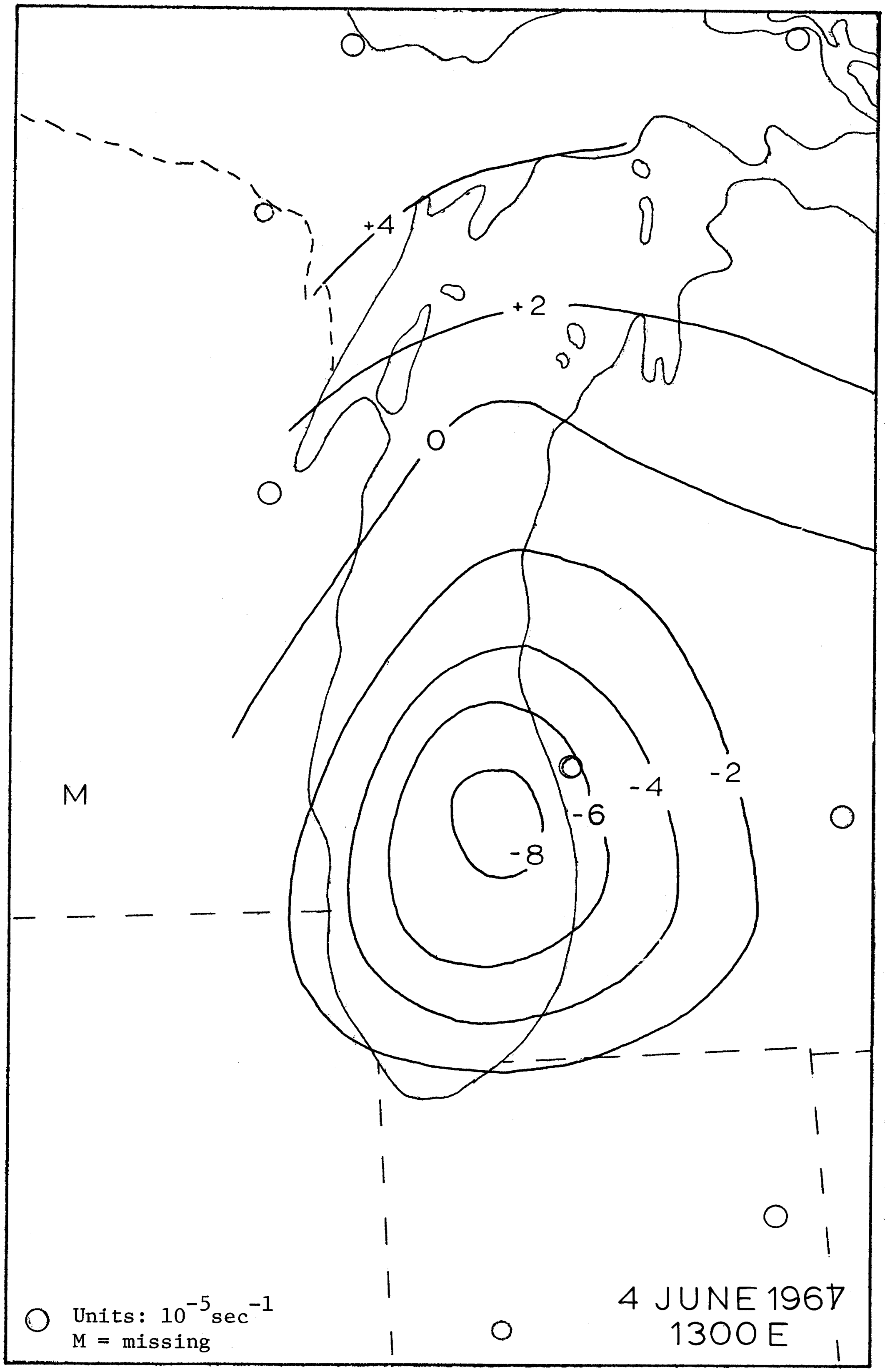

FIG. A28. 610 m absolute vorticity - 4 June 1967 at 1300E.

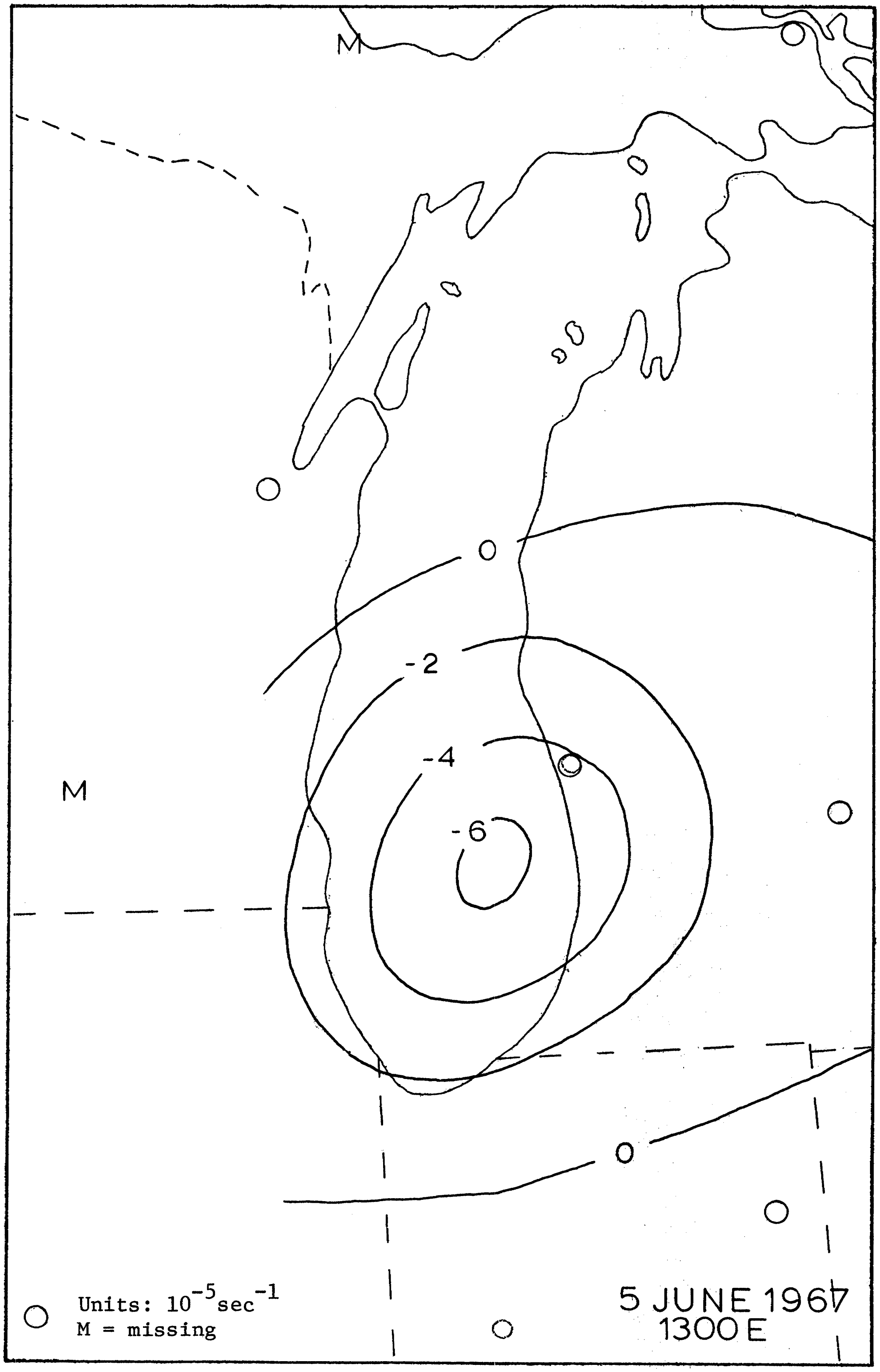

FIG. A29. 610 m absolute vorticity - 5 June 1967 at 1300E.

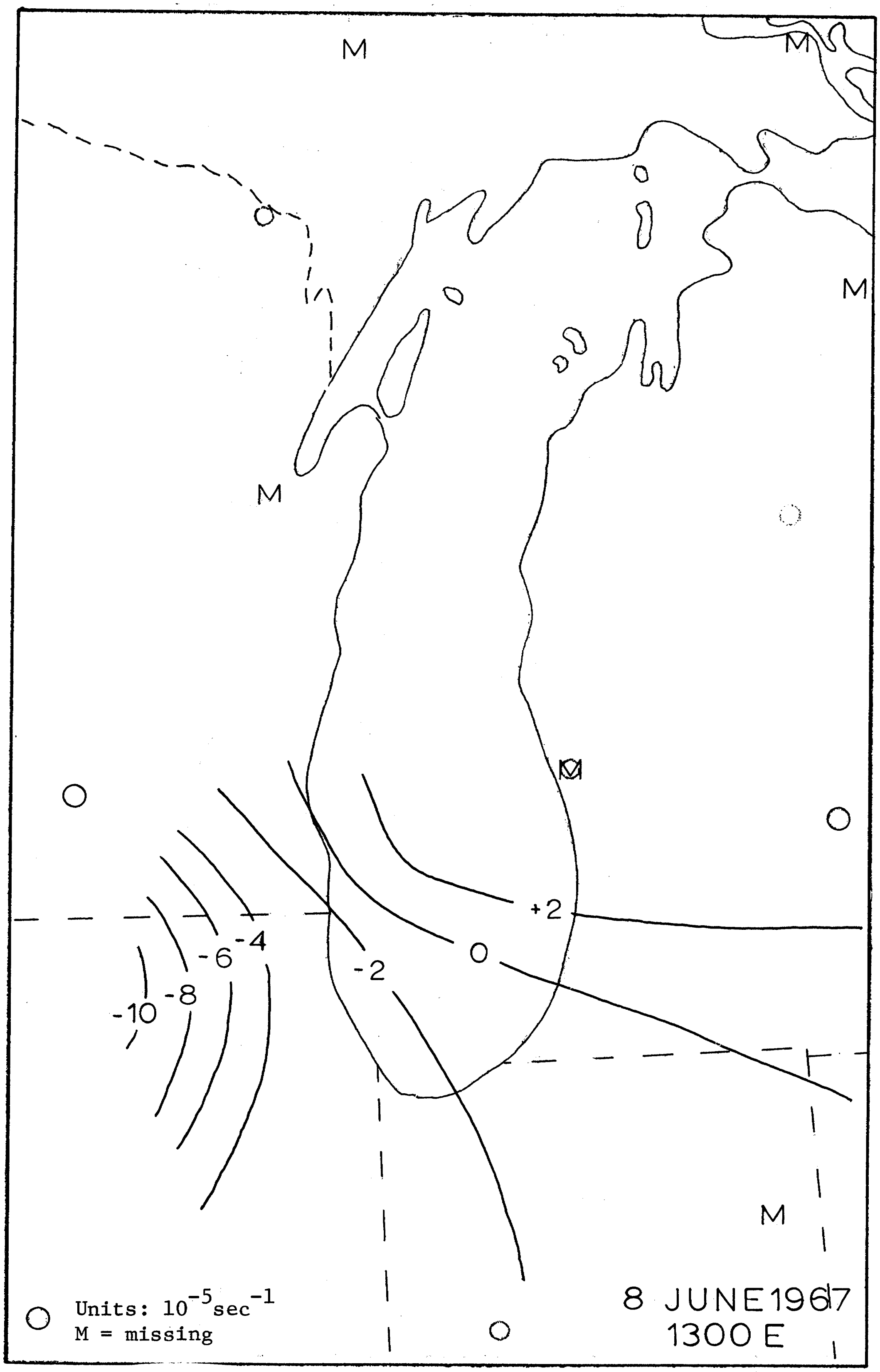

FIG. A30. 610 m absolute vorticity - 8 June 1967 at 1300E.

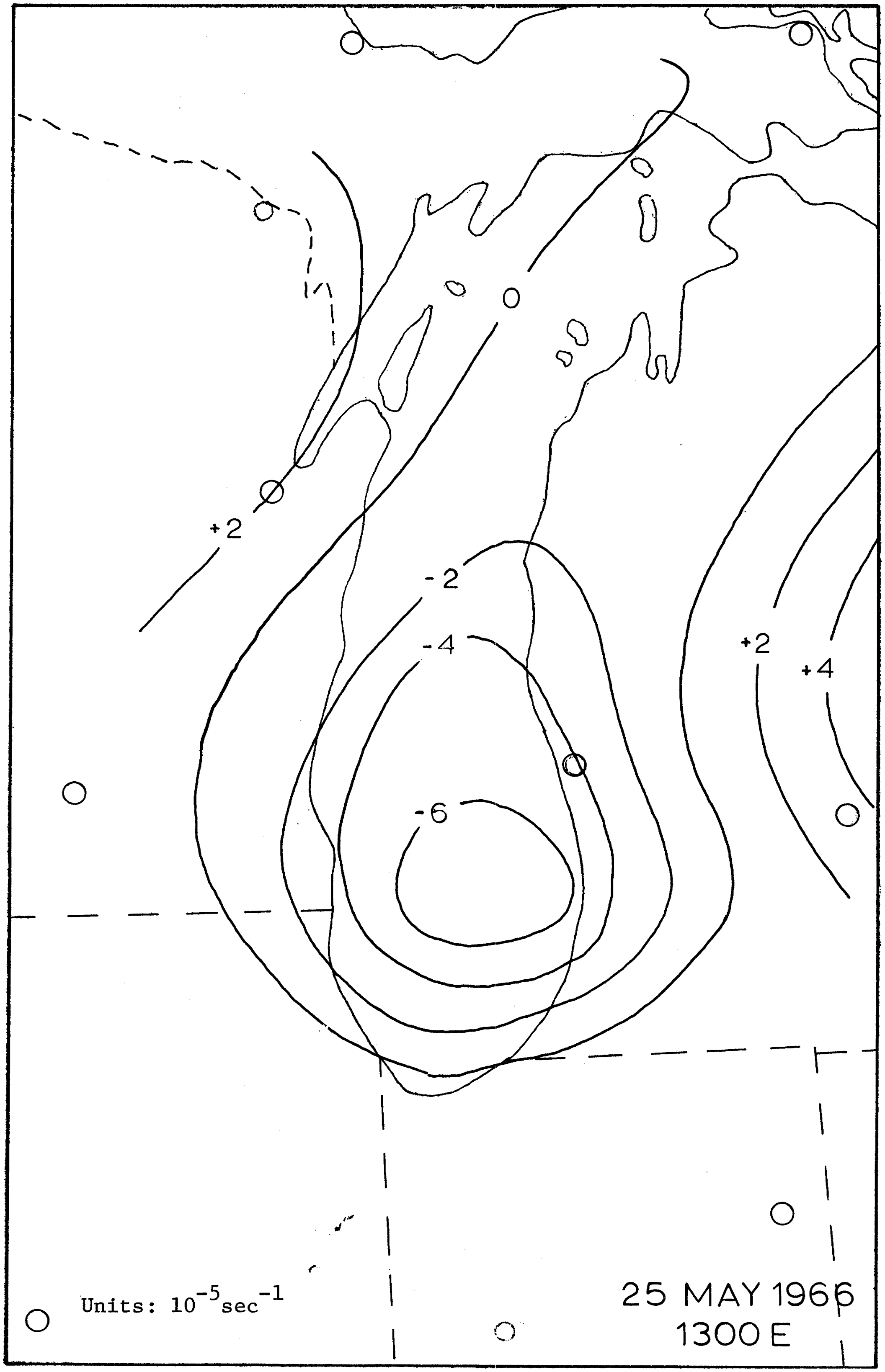

FIG. A31. 610 m absolute vorticity - 25 May 1966 at 1300E.

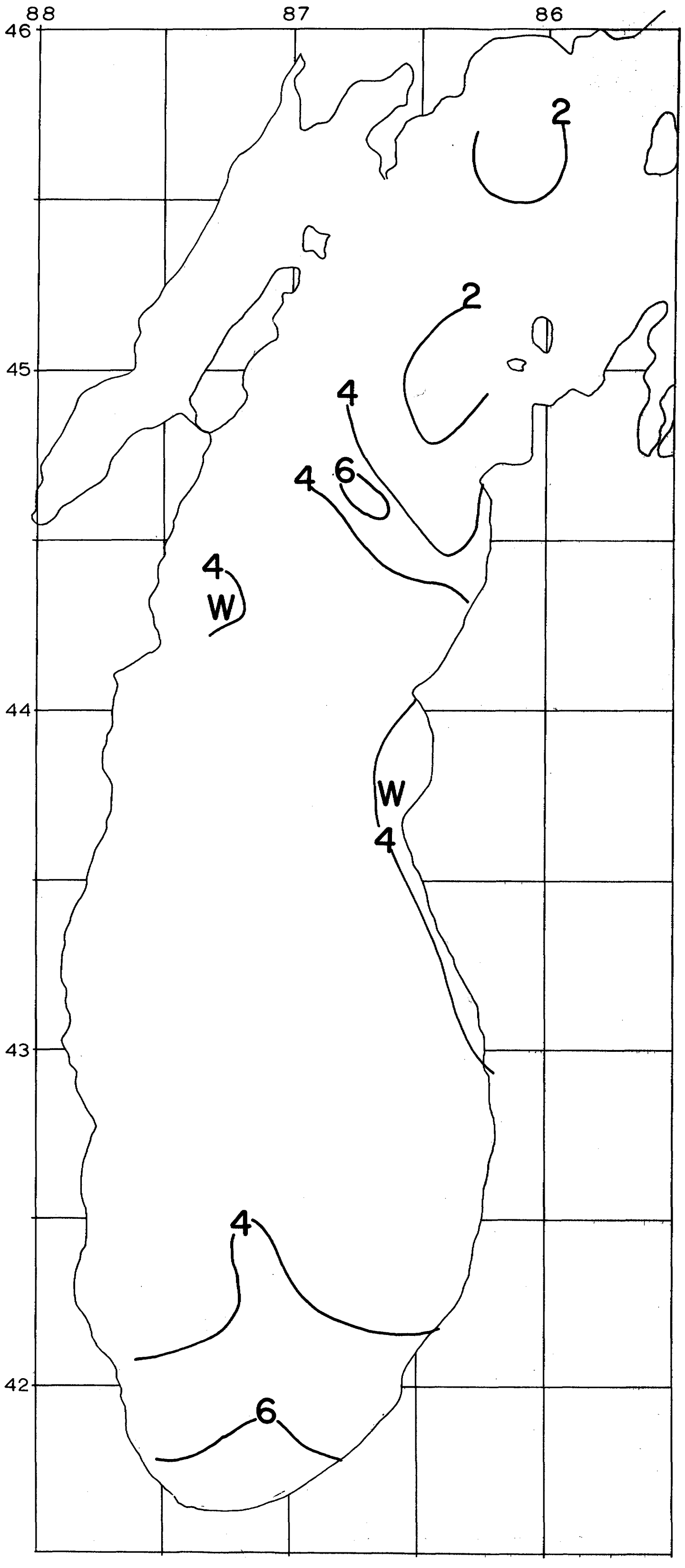

FIG. A32. Average surface water temperature (°C) for the week: 30 April - 6 May 1966. W = warmer waters.

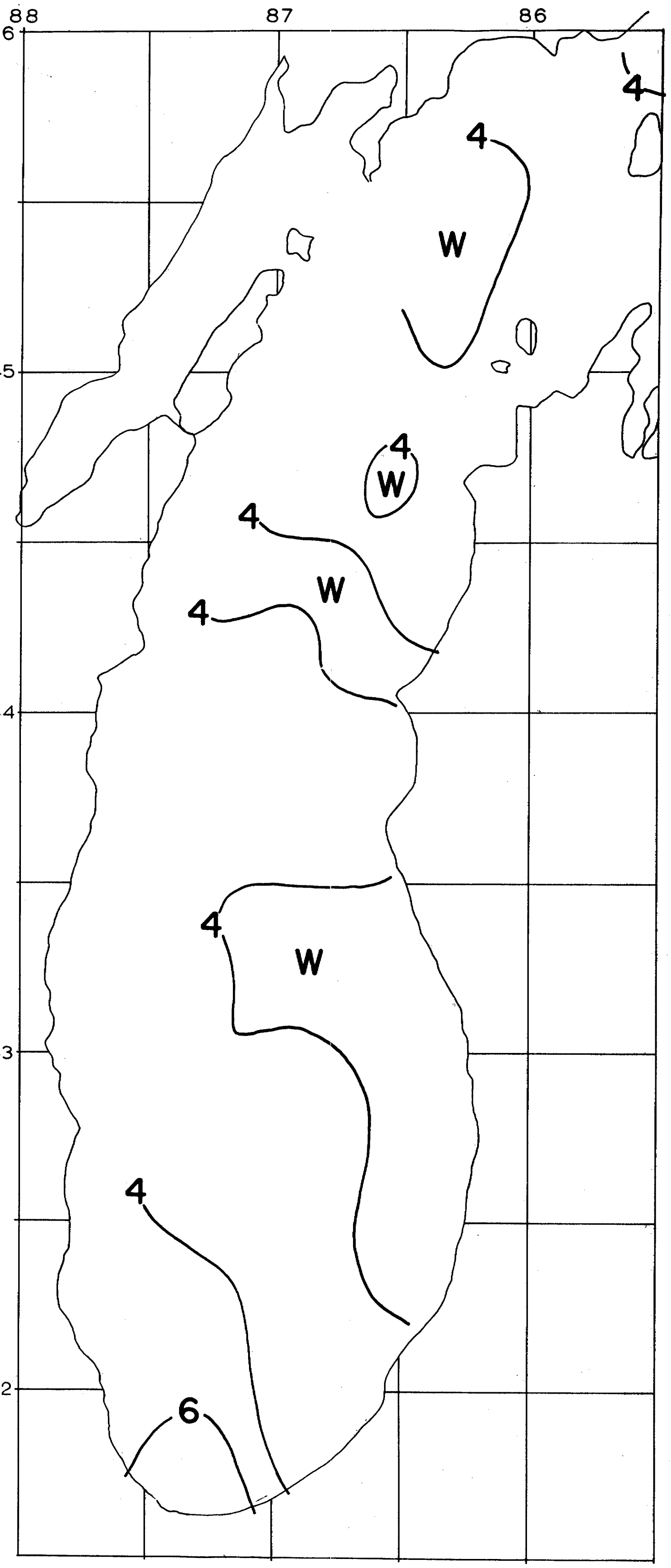

FIG. A33. Average surface water temperature (°C) for the week: 7 - 13 May 1966.

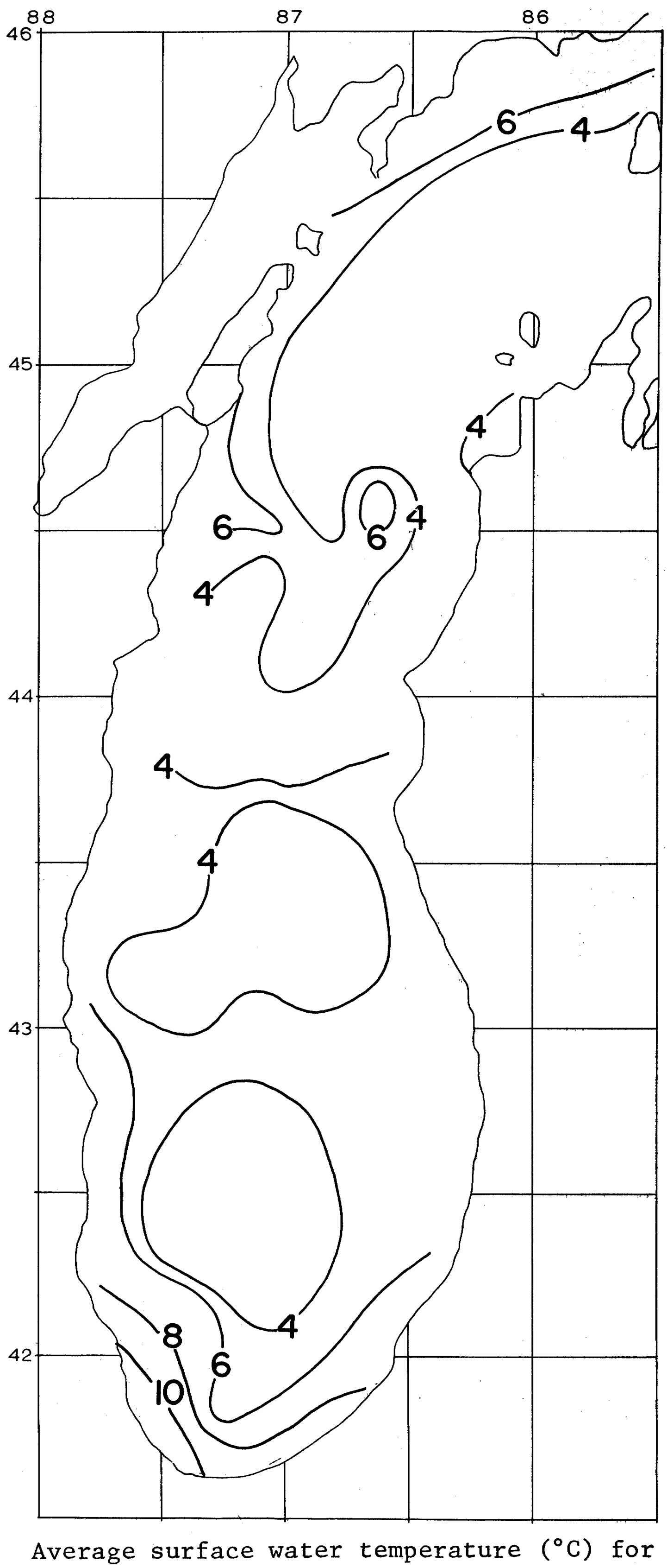

FIG. A34. Average surface water temperature (°C) for the week: 14 - 20 May 1966.

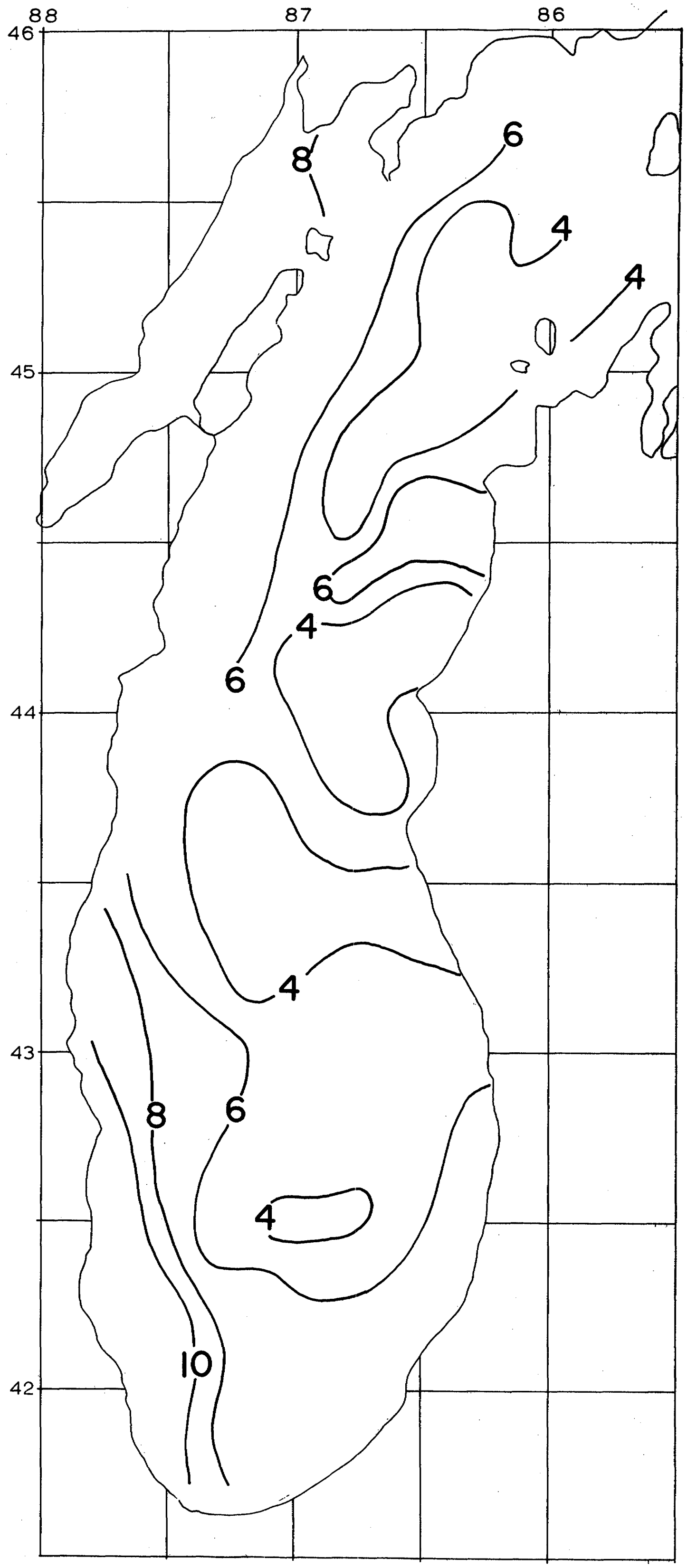

FIG. A35. Average surface water temperature (°C) for the week: 21 - 27 May 1966.

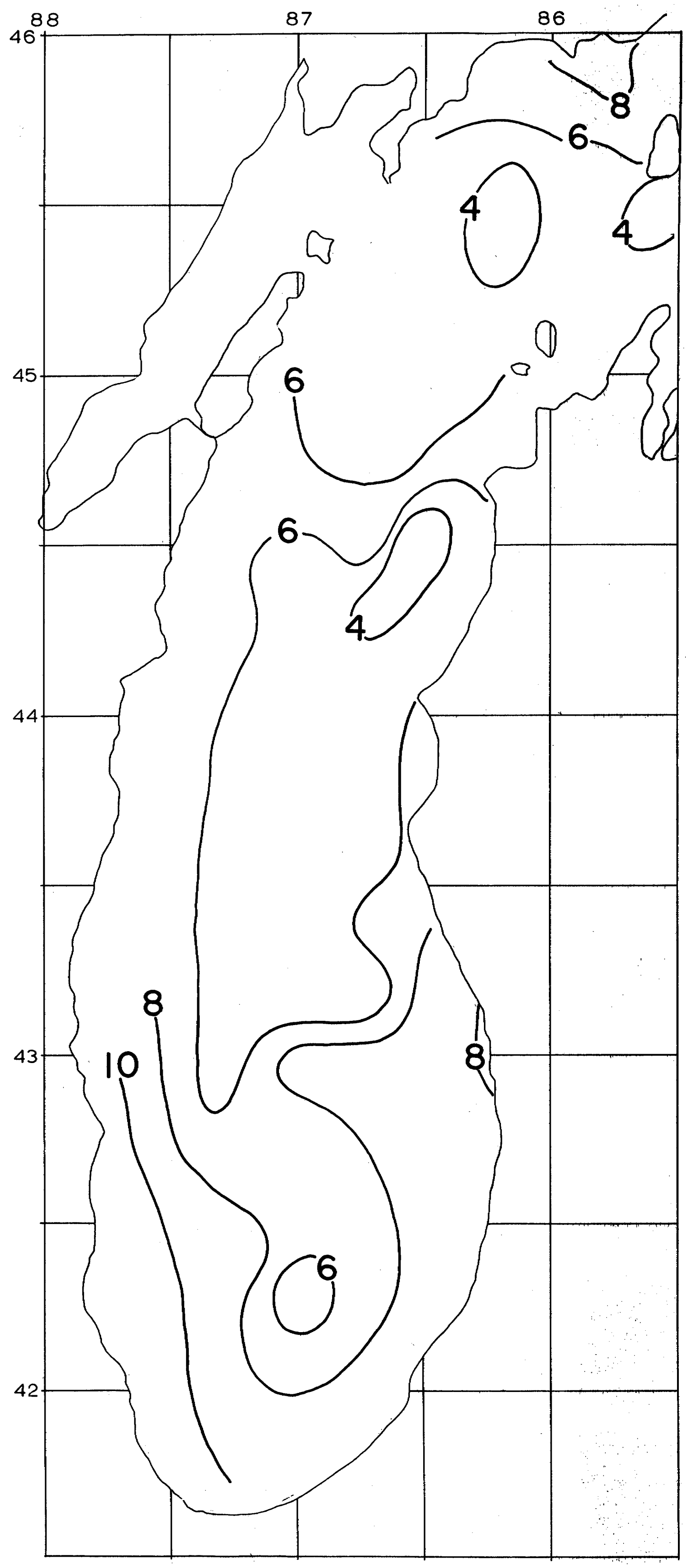

FIG. A36. Average surface water temperature (°C) for the week: 28 May - 3 June 1966.

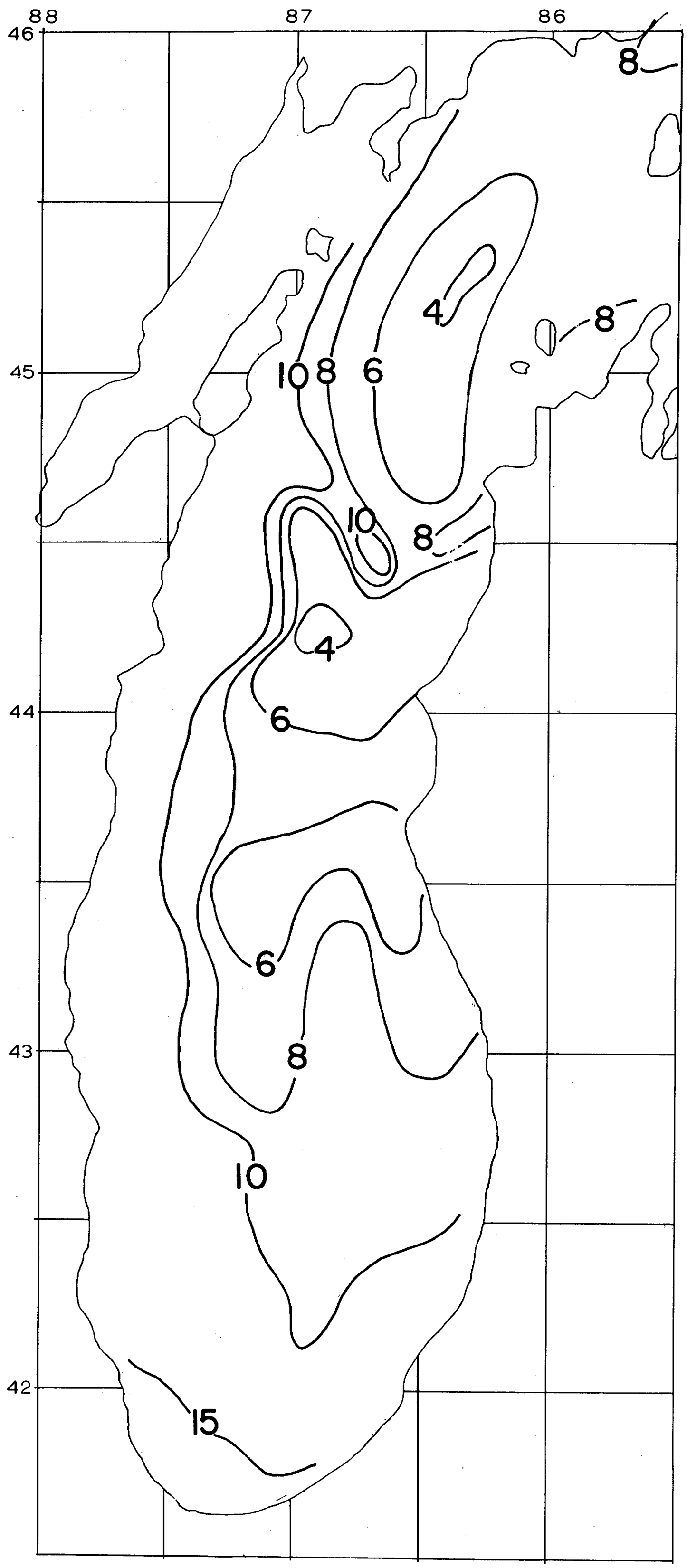

FIG. A37. Average surface water temperature (°C) for the week: 4 - 10 June 1966.

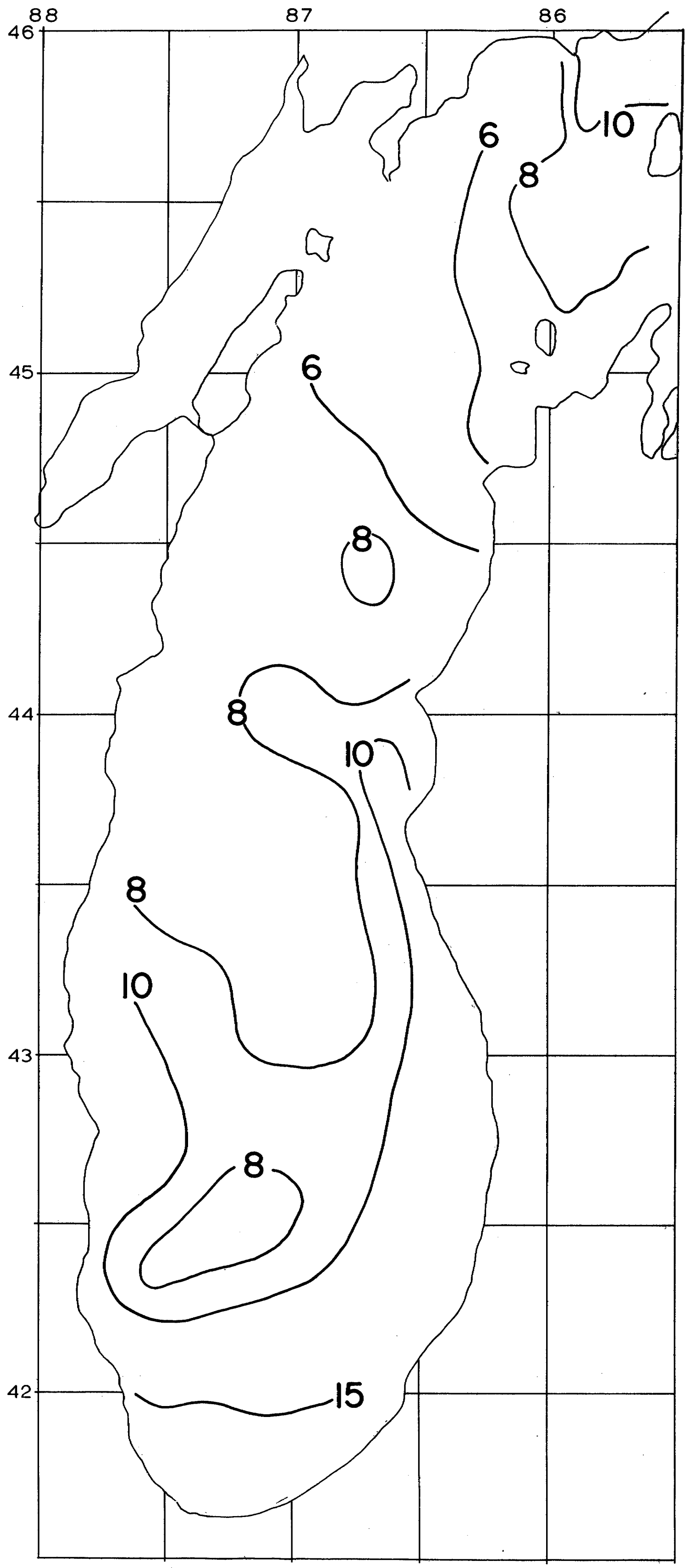

FIG. A38. Average surface water temperature (°C) for the week: 11 - 17 June 1966.

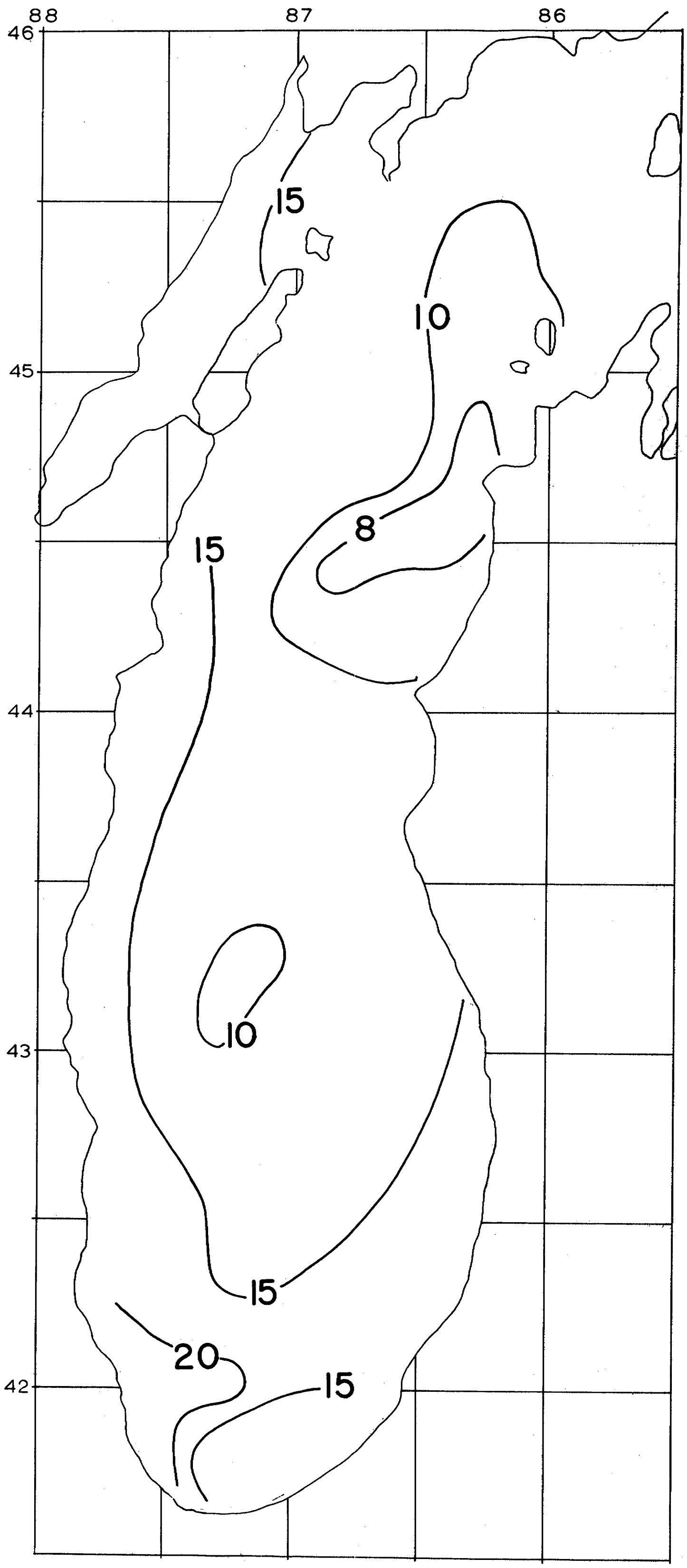

FIG. A39. Average surface water temperature (°C) for the week: 18 - 24 June 1966.

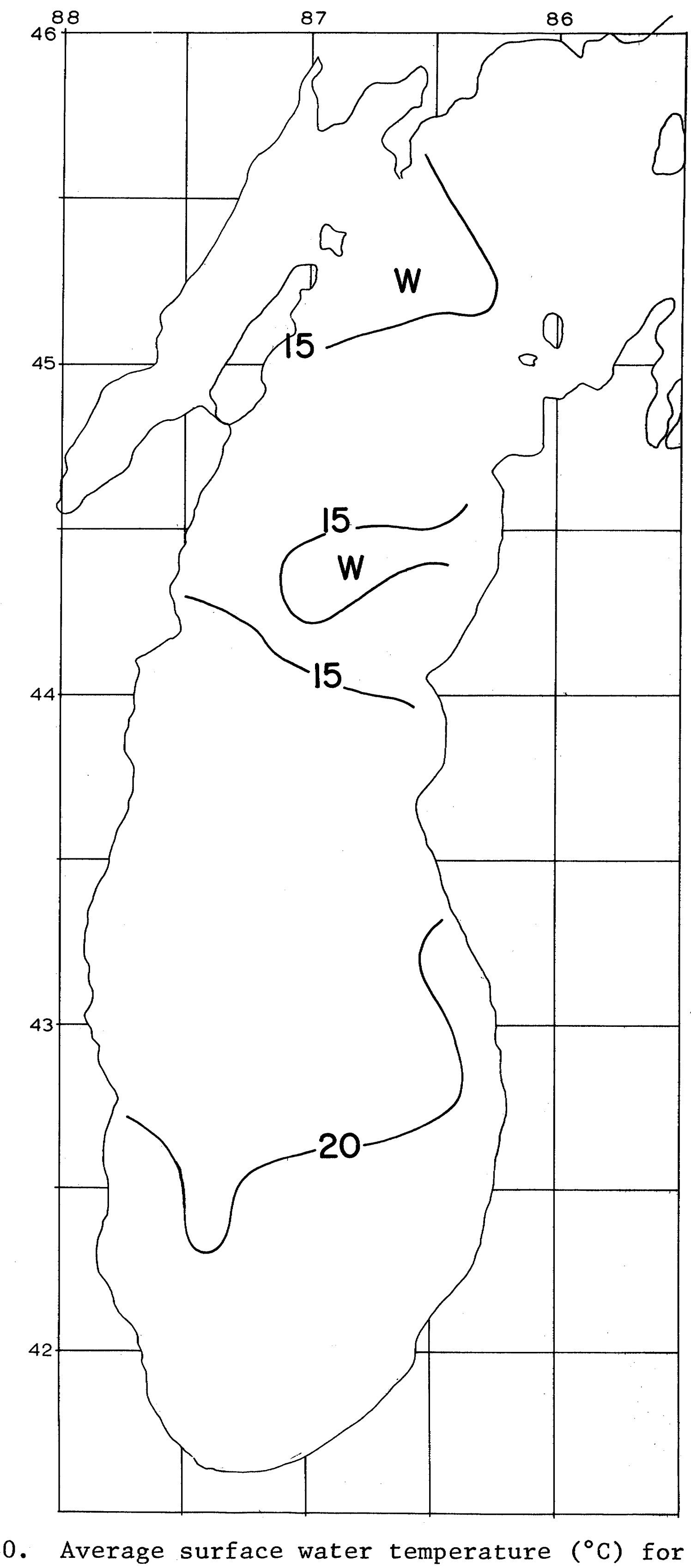

FIG. A40. Average surface water temperature (°C) for the week: 25 June - 1 July 1966.

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